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CONTACT AND ENDOMORPHIC PHENOMENA ASSOCIATED WITH A PART OF THE IDAHO BATHOLITH

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ABSTRACT

The intrusion of the Idaho batholith in the Paradise Ridge area of north Idaho was accompanied by clearly defined contact and endomorphic phenomena. Hydrothermal solutions emanating from deeper parts of the consolidating magma changed a part of the early, chilled marginal hornblende-biotite diorite of the batholith to a less calcic biotite-quartz diorite and some of the quartzite above to biotite-quartz diorite and to a quartz-biotite-diorite gneiss. In a part of the area the emanations were enriched in potash and these, permeating through the gneissic rock, changed much of it to an augen gneiss with microcline porphyroblasts and the adjacent biotite-quartz diorite to granodiorite. Toward the end of the period of igneous activity the emanations were highly silicic and introduced exceptionally large amounts of quartz into all the rocks, even causing silicification of the invaded quartzite. Minerals added to the rocks by hydrothermal replacements include biotite, andesine, microcline, quartz, sphene, zircon, apatite, magnetite, garnet, allanite, zoisite, epidote, muscovite, sericite, and calcite.

As consolidation progressed downward and hydrothermal solutions were drawn off from deeper levels, the solutions moved upward along fractures in the already endomorphosed rock and by replacement changed the bordering rock to coarse but variable-grained bodies of biotite pegmatite. One of these bodies is extraordinarily rich in garnet. As the solutions became more silicic, fissures were filled with quartz, or the wall rock was replaced by quartz, and bodies of silexite were formed.

This study shows the importance of igneous emanations in the formation of quartz-feldspathic gneisses, augen gneisses, and pegmatites, and the extent to which hydrothermal replacements enter into their origin. It is suggested that such phenomena may be widespread in the crystalline schists of extensive metamorphic terranes.

INTRODUCTION

On Paradise Ridge in Latah County, Idaho, there is a group of rocks which show the effects of certain exomorphic and endomorphic processes associated with and taking place within and about the borders of the Idaho batholith. The metamorphism has caused progressive changes in both the intruding and the intruded rock by the addition of material from hydrothermal solutions given off during consolidation of the invading magma. Some of the changes are not unlike those observed about the Cassia batholith in the south part of the state, where contact effects are represented by the impregnation of "igneous material" into quartzite in such a manner as to make it resemble granitic rock.[1] However, on Paradise Ridge impregnation of "igneous" material under differential stress made gneissic rather than granitoid rock out of the quartzite, and the phenomena are such as to bear on the problem of the origin of gneissic rocks in extensive metamorphic terranes, from the point of view that such rocks are possible products of igneous emanations and hydrothermal replacements. This study also provides an explanation for the origin of augen gneisses other than by dynamic metamorphism of pre-existing porphyritic granitic rocks.

Paradise Ridge is the name of a part of a more or less continuous ridge which extends in a general westerly direction from the Clearwater Mountains (a part of the northern Rockies) far out upon the Columbia Plateau. The ridge comes to an end at Bald Ridge Butte just over the state line in Washington, where its geology has been discussed in some detail by Hoffman.[2] Paradise Ridge is about $3\frac{1}{2}$ miles long and rises to an altitude of almost 3,500 feet above sea-level, or nearly 1,000 feet above the surface of the plateau. The part of the ridge which was mapped and studied in detail (Fig. 1) covers about $2\frac{1}{2}$ square miles and lies about 9 miles east-northeast of the Bald Ridge Butte area, or about 3 miles southeast of Moscow, Idaho.

[1] A. L. Anderson, "Contact Phenomena Associated with the Cassia Batholith, Idaho," *Jour. Geol.*, Vol. XLII (1934), pp. 376–92.

[2] M. G. Hoffman, "The Geology of Bald Ridge Butte, Washington," *Jour. Geol.*, Vol. XL (1932), pp. 634–50.

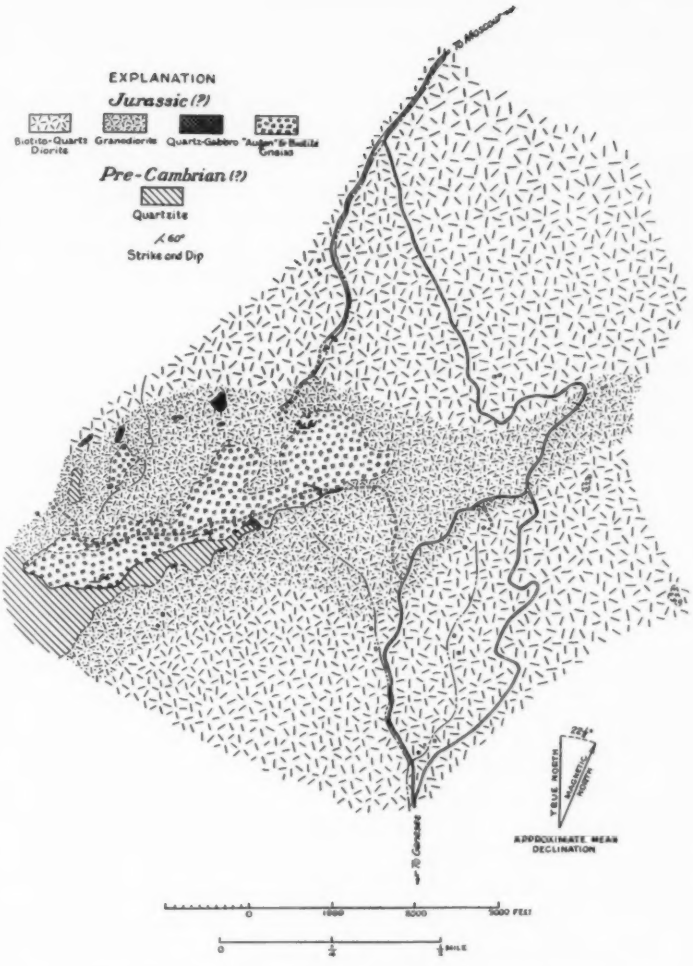

FIG. 1.—Geologic sketch map of a part of Paradise Ridge, Latah County, Idaho

GEOLOGIC FEATURES

The area contains a variety of igneous and metamorphic rocks; namely, quartzite, quartz gabbro, hornblende-biotite diorite, biotite-quartz diorite, quartz-biotite-diorite gneiss, augen gneiss, granodiorite, pegmatite, and silexite.[3] The quartzite is the oldest rock and may be pre-Cambrian, or, as Hoffman suggests, early Paleozoic.[4] It has been intruded by a part of the Idaho batholith (late Jurassic or Cretaceous);[5] and a part of it has been changed to gneiss, which is also in part an augen gneiss. Erosion has removed the gneissic and quartzitic covering from much of the area; but, locally, removal of the batholithic rock has just begun.

The quartzite and gneissic rocks occur as narrow bands elongate parallel to the ridge. Both are more resistant to erosion than the granitic rocks and occur for the most part along the crest and upper slopes of the ridge. The quartzite, being more resistant to erosion than the gneissic rocks, forms a higher part of the ridge than the latter. However, a few isolated remnants of each appear on some of the spur ridges at somewhat lower levels. The quartzite band enlarges westward and makes up an increasingly larger part of the ridge. All primary structural features of the quartzite have been obliterated by recrystallization. The bedding has been wholly destroyed or is so poorly defined that strikes and dips are not determinable. The quartzite contains many joints and in places has been sheared and fractured. The most extensive shearing and fracturing has been confined to a distinct zone extending for about a mile along the crest of the ridge. Localization of the gneissic rock along this zone of weakness has almost destroyed the original sheared and fractured quartzite. The band of gneiss now appears to lie alongside and to extend beyond the east end of the quartzite (Fig. 1). Foliation in the gneiss is well marked. The strike of the banding is nearly due east or but a few degrees either north or south of east. The dip

[3] Not all the varieties are in bodies large enough to show on the map. Rock designations have been simplified in the map legend and some varieties combined in a simple pattern.

[4] *Op. cit.*, p. 637.

[5] C. P. Ross, "Mesozoic and Tertiary Granitic Rocks in Idaho," *Jour. Geol.*, Vol. XXXVI (1928), pp. 686–92.

is, with few exceptions, to the north at angles ranging from 30° to 60°.

Much of the batholithic rock is a biotite-quartz diorite, but locally, it grades into granodiorite. These rocks can be separated only in a general way in the field, but detailed mapping aided by microscopic study shows that much of the granodiorite forms an irregular envelope around the mass of quartzite and gneiss (Fig. 1) and that it is more or less transitional from biotite-quartz diorite into augen gneiss. Evidence of some primary foliation in these rocks is suggested by a crude parallel arrangement of biotite crystals and by an approximate east-west alignment of scattered inclusions of older rock. These inclusions consist of narrow, well-banded, dark, fine-grained fragments of hornblende-biotite diorite, here and there drawn into schlieren-like shapes. Most of the dark fragments are small and average from less than 1 to 3 feet thick, but some are as much as 15 feet thick. The inclusions may be rounded or angular and appear to be sharply bordered or to shade more or less gradually into the biotite-quartz diorite. Because a thin mantle of soil conceals the bedrock over much of the area, the dark inclusions are rarely observed except in quarries and road cuts. Their abundance in such exposures suggests that they may be widely distributed. In some places to the west of the map area, the dark-colored, fine-grained hornblende-biotite diorite, which makes up the inclusions, forms a chilled marginal zone on the batholith and is commonly cut by apophyses of the lighter-colored and coarser-grained biotite-quartz diorite. The inclusions in the biotite-quartz diorite in the map area apparently represent incorporated masses of the shattered, earlier, chilled, and consolidated border rock.

Pegmatites are abundant in the biotite-quartz diorite and granodiorite. Few are found in the quartzite and gneiss. Most of the pegmatite dikes are small and only two could be traced for more than 100 feet. The thickness of most of them ranges from a fraction of an inch to 1 or 2 feet, exceptionally to 10 or 15 feet. Few of the pegmatites have sharply defined walls. Most of them pass more or less gradually into the enclosing country rock. Near some of the pegmatites the quartz diorite has streaks or thin seams of biotite aligned along the faintly developed foliation. There are also irregular and

podlike masses of quartz and of feldspar in some of the granitic rock. The position of the pegmatites is controlled by joints and by fractures in the host rock.

Dikes and pod-shaped masses of silexite are fairly common, particularly in the gneissic rocks. Some of these appear to show transition into pegmatites.

The quartz gabbro is more difficult to place. Bodies of it are small, rounded, or more or less elliptical, and appear to be confined mainly to the granodiorite. The bodies are so poorly exposed that it is impossible to determine whether or not they actually intrude the granodiorite. Hoffman likewise had difficulty in determining actual relationships in the Bald Butte area. He described the bodies as dikelike and reported them as apparently intruding aplite but as being cut by pegmatites and quartz dikes.[6] In general, exposures are not more than a few feet long or wide; many are less than a foot. Weathered surfaces of the rock appear bedded, and they so closely resemble an altered fine-grained quartzite that they were regarded as such until examined under the microscope. Microscopic study shows that the gabbro has been metamorphosed and, therefore, must have been intruded early, but how early is an unsolved problem.

PETROGRAPHY

QUARTZITE

Much of the quartzite is composed of a white, rather coarsely granular rock. In places the quartz is somewhat turbid; but in other places it is distinctly glassy like that in pegmatites or in high-temperature hydrothermal veins. As primary sedimentary structures have been largely destroyed, exposures of the quartzite are not everywhere easy to distinguish from those of silexite. Although notably massive and uniform in composition, the quartzite in places shows evidence of shearing and is broken by tiny subparallel to parallel fractures which tend to give it a more or less distinct foliation. In some of the minutely fractured rock are visible grains of muscovite and in some rudely aligned grains of feldspar and biotite. As the mica and feldspar increase in abundance the quartzite resembles gneiss. This change by impregnation with feldspar and mica

[6] *Op. cit.*, pp. 640–41.

becomes more noticeable near the mass of quartz-biotite-diorite gneiss. By further increase in the feldspathic constituents the rock actually changes to gneiss.

Microscopic studies show that the quartzite has been thoroughly recrystallized with all traces of former sand grains completely obliterated. In the recrystallized rock the quartz grains apparently have coalesced into larger masses that show pronounced undulatory extinction and have wavy, lacelike borders. As a result the quartzite appears as a coarse mosaic of interlocking grains, except that the mosaics have been more or less extensively penetrated by irregular quartz veinlets which are easily discerned in thin section (Fig. 2, *A*) but not in the hand specimen. In some places the quartz veinlets form such a large part of the rock that the original grains of the quartzite are preserved only as residual remnants which are retained as engulfed inclusions. These veinlets penetrate unevenly into the grains of the quartzite and also cut directly through them irrespective of fracture or grain boundaries (Fig. 2, *A*). Many of the penetrating grains are typically lobate in form. Because this phenomenon is observed so widely through the body, the quartzite appears to have been more or less extensively silicified, which probably accounts for its resemblance to vein quartz in the outcrop.

Where the quartzite has been sheared there are minerals which lie along the fractures and grain contacts and which also extend into and through the quartz grains as if by replacement. These minerals include andesine (about Ab^{65}), biotite, quartz, and variable but generally small amounts of microcline, muscovite, zircon, apatite, garnet, sphene, and magnetite. The biotite crystals commonly are oriented with their longer dimensions parallel to the cleavage, and, where the biotite becomes relatively abundant, the rock is schistose. The feldspars and the biotite have imperfect crystal outlines and occur as irregularly shaped grains penetrating and holding as inclusions remnants of the original quartz. Both the feldspar and biotite grains have been embayed and invaded by lobate grains and veinlets of quartz such as penetrate the main mass of the quartzite (Fig. 2, *B*). Andesine partly replaced by the quartz is not uncommonly altered to albite, zoisite, and sericite. The sphene, zircon, apatite, and magnetite favor a distribution along fractures but also appear along con-

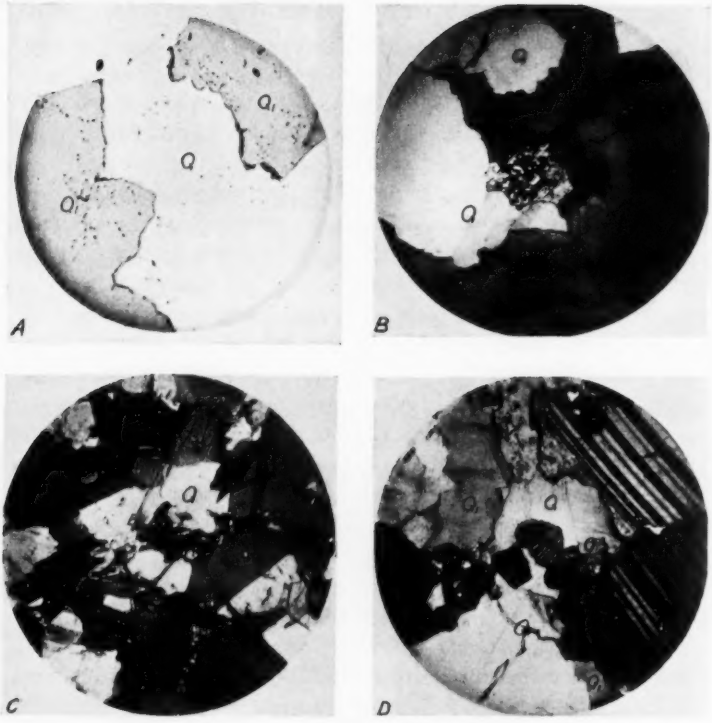

FIG. 2.—*A*, Photomicrograph of the recrystallized quartzite showing a seam of hydrothermal quartz (*Q*) cutting directly through and replacing an original quartz grain (Q_1). Mag. 24×. *B*, Photomicrograph of the quartzite which contains a grain of plagioclase embedded in the quartz by replacement. The plagioclase and quartzite are invaded and partly replaced by lobes of younger quartz (*Q*). Note how the end of the plagioclase grain has been cut off and unevenly embayed by the quartz lobe. Mag. 21×. *C*, Photomicrograph of the marginal hornblende-biotite diorite showing the mineral relationships which were produced by crystallization during consolidation of the early chilled magma. The picture shows a grain of quartz (*Q*) which was added to the rock and which embays and replaces biotite (*B*) and hornblende (*H*). Mag. 21×. *D*, Photomicrograph of an inclusion of marginal diorite engulfed in granodiorite which shows a seam of quartz (*Q*) cutting and embaying unevenly crystals of andesine and grains of an earlier quartz (Q_1). Mag. 21×.

tacts of the invading quartz and along borders of the other minerals. Muscovite grains likewise appear along grain borders and penetrate both the early and the late invading quartz. Some muscovite occurs far out in the main body of silicified quartzite

HORNBLENDE-BIOTITE DIORITE

The hornblende-biotite diorite, which comprises the numerous inclusions of the earlier, quickly chilled border facies of the batholith, is a dark gray, rather fine-grained rock with the average grains about 1 millimeter in diameter. That much of the rock is somewhat gneissic is more evident in the thin section than in the hand specimen. In most places the contact between the inclusions and the biotite-quartz diorite and granodiorite host is sharply defined, but in some places there is a marked gradation from one into the other.

Hornblende and biotite are each plainly visible in the hand specimen and together make up about 15–20 per cent of the rock. They show a rather irregular distribution, and either one may exceed the other in any one place. On an average they are perhaps about equally abundant. Andesine is the leading mineral and normally composes about 70 per cent of the rock. The other minerals include numerous needles of apatite; large grains of epidote; small crystals of zircon, magnetite, and sphene; and ordinarily very subordinate amounts of quartz. There is also some chlorite, sericite, and calcite.

The andesine grains have no unusual features. Most of them show little sign of alteration; those which do are partly altered to minute grains of white mica, zoisite, and calcite. However, some of the andesine crystals are penetrated by large grains of epidote which commonly have tonguelike offshoots extending along and across cleavages. Some of the epidote also impregnates and is more or less pseudomorphous after hornblende. Biotite may or may not be intimately associated with the hornblende. Where it is, it may penetrate or form a broken mantle on the hornblende. The biotite also may penetrate the andesine in such a way as to suggest that it has had an added growth and has in part replaced the andesine. A part of the biotite has altered to chlorite. In most places the quartz appears to form not much more than 5 per cent of the rock; but in some places it is present in abundance. Much of the quartz is lobate in

form and tends to penetrate the andesine and dark minerals (Fig. 2, *C*) and to replace them in the same way as does the quartz which invades the quartzite. Accessory minerals show a preference for the dark minerals and are commonly included in them; but some of the accessory grains are aligned along quartz-andesine borders and along cleavages in andesine.

Some of the hornblende-biotite diorite which occurs as inclusions in the biotite-quartz diorite seems to have been permeated by the containing host. In such occurrences the hornblende and epidote of the inclusion appear to have been lost and their places taken by increased amounts of biotite, but the inner part of the inclusion is not unlike the sharply delineated ones described above. In the actual transition zone the andesine and the biotite are extensively penetrated by quartz and tend to assume the appearance of remnant inclusions (Fig. 2, *D*). Where the change is complete, the andesine of the diorite has been replaced by a somewhat less calcic one (the one typical of the biotite-quartz diorite or granodiorite). Some of the less calcic andesine in turn has been penetrated, corroded, and in part incorporated as residual inclusions in orthoclase. Both feldspars may be veined and penetrated by lobes of quartz; and the original diorite locally shows a more or less complete change into granodiorite. Except as remnant inclusions, much of the biotite has been lost during the transformation. Muscovite may appear along grain borders and in fractures in the feldspars. Accessories have the usual relations to grain borders and cleavages.

BIOTITE-QUARTZ DIORITE

The biotite-quartz diorite, the most widespread of the rocks on Paradise Ridge, is the more or less typical marginal rock along the northwest border of the Idaho batholith.[7] It is light to moderately dark gray, medium to fairly coarse-grained, the average grains ranging from 3 to 5 millimeters. The rock is commonly somewhat gneissic and in places is streaked by thin lenticular seams of biotite. It is not easy to distinguish some of this rock from the near-by biotite-quartz-diorite gneiss.

The mineral composition is essentially andesine, quartz, and bio-

[7] A. L. Anderson, "The Geology and Mineral Resources of the Region about Orofino, Idaho," *Idaho Bur. Mines and Geol. Pamph. 34* (1930), pp. 17–19.

tite. The andesine (about Ab^{65}) appears to have the same composition as that in the "granitized" quartzite. It forms about 35–65 per cent of the rock with the usual limits between 50 and 60 per cent. The quartz content ranges from about 25–35 per cent, but an upper limit of 60 per cent has been observed. Biotite forms from 5 to 10 per cent of the rock, but exceptionally much higher amounts have been noted. In places muscovite is relatively abundant. The accessory minerals include orthoclase, apatite, zircon, magnetite, and a very little garnet and sphene. Other minerals are sericitic mica, chlorite, and calcite.

The grains of andesine are ordinarily clear and show little evidence of alteration except for scattered irregular patches of secondary white mica distributed especially along cleavages. However, some of the grains hold phantom-like outlines of a somewhat more calcic plagioclase (Fig. 3, *A*), and others, inclusions of rounded, corroded quartz grains (Fig. 3, *B*) and remnants of biotite grains. Some of the biotite grains lie at the margin of the andesine crystals and appear as though the andesine crystals had penetrated into them. On the other hand, there are also biotite grains of another habit which lie along fractures in the andesine and penetrate the andesine from crystal margins. In the rock, which is streaked by long, thin seams of biotite, the andesine is penetrated and replaced extensively by the biotite. The dual habit of the biotite suggests that it is present in two generations, one older and one younger than the andesine.

The quartz also has more than one habit. In addition to that quartz which occurs as residual inclusions in the andesine, there is a much larger amount which forms lobate grains and irregular veinlets that penetrate the andesine and biotite crystals. These penetrating grains and irregular veinlets give the andesine and biotite crystals very uneven outlines. In places the andesine and biotite grains are enclosed wholly in the quartz. Some of the lobate quartz and quartz veinlets contain rounded and irregular grains of quartz which may originally have filled in between grains of andesine. Some of the lobate grains also penetrate other lobate grains (as in Fig. 2, *D*) and contain parts of them as inclusions. These relations suggest that there are at least two and probably three different stages of quartz deposition exclusive of the early quartz, which occurs as inclusions in

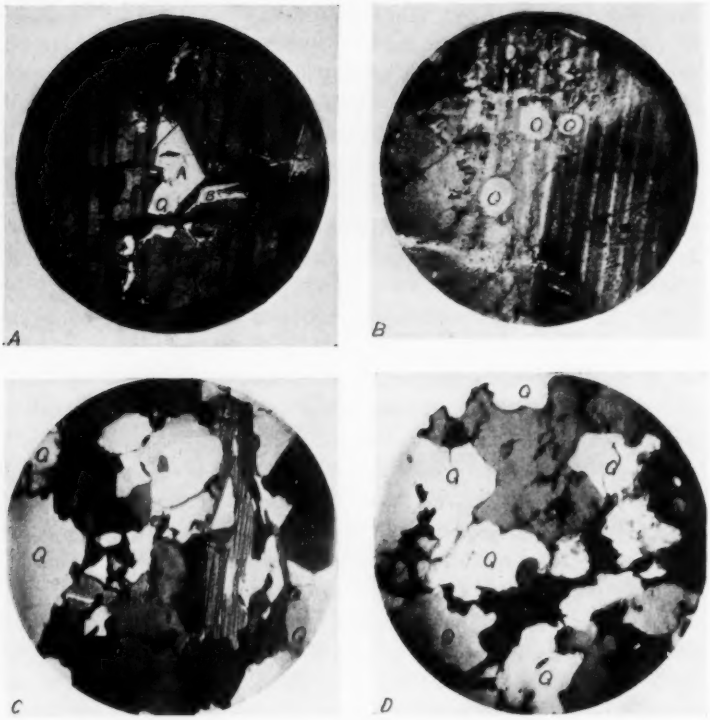

FIG. 3.—*A*, Photomicrograph of the biotite-quartz diorite showing an andesine crystal with remnants of an earlier, more calcic andesine (*A*), quartz (*Q*), and biotite (*B*), inherited from the marginal diorite, which the biotite-quartz diorite replaces. Mag. 24×. *B*, Photomicrograph of the biotite-quartz diorite showing an andesine crystal with rounded, remnant inclusions of quartz (*Q*) inherited from the quartzite which the biotite-quartz diorite replaces. Mag. 24×. *C*, Photomicrograph of the quartz-biotite-diorite gneiss with a remnant of micaceous quartzite which largely fills the field of view. Note the approximate parallelism of the biotite which replaces the quartzite, and note, too, the presence of lobate quartz grains (*Q*) which invade and replace the biotite and quartz of the quartzite remnant. Mag. 21×. *D*, Photomicrograph of the quartz-biotite-diorite gneiss showing lobate quartz (*Q*) penetrating extensively into the plagioclase and biotite of the gneiss. Mag. 21×.

the andesine. The abundance of the late lobate grains and seams of quartz has given the rock its exceptionally high quartz content. The relations of the grains and veinlets suggest widespread silicification of the rock after it had become partly or wholly solid.

Orthoclase, observed in but a few sections, tends to penetrate and corrode the andesine. Some of the andesine so penetrated contains myrmekitic-like blebs of quartz. Like the andesine the orthoclase is more or less penetrated by quartz lobes. Muscovite is relatively conspicuous in places as fairly large grains aligned along cleavages in the feldspars, also along grain borders and fractures. Zircon, apatite, magnetite, and garnet likewise favor a distribution along grain borders, cleavages, and fractures. Their distributions and associations suggest that they are about the same age as the late quartz. Chlorite is the common alteration product of the biotite. However, some of the muscovite is pseudomorphous after biotite. Calcite is present as late grains.

QUARTZ-BIOTITE-DIORITE GNEISS

The quartz-biotite-diorite gneiss occurs as irregular masses enclosed in augen gneiss and as small scattered bodies in the granodiorite and biotite-quartz diorite. Much of the rock is conspicuously foliated and is a typically banded gneiss with biotite in thin parallel bands, alternating with bands of quartz and feldspar. Its grain is medium and its color rather dark, but not much darker than the biotite-quartz diorite. In places the feldspathic bands tend to assume lenticular shapes; and, as the lentils increase in prominence, the rock changes to augen gneiss.

Much of the quartz-biotite-diorite gneiss has the approximate composition of quartz diorite, but some is near granodiorite. Because of marked variations in composition of the gneiss in short distances, no rock can be described as typical; but an average specimen may show about 50 per cent andesine (Ab^{65}), 0–15 per cent microcline, 20–30 per cent quartz, and about 15 per cent biotite. Accessory and secondary minerals not estimated in the total are zircon, apatite, magnetite, rutile, sphene, garnet, allanite, epidote, zoisite, and finely divided white mica.

Although the minerals of the rock are about the same as those in

the biotite-quartz diorite and the granodiorite, the textural relations in all but a few places are very different. The foliation reflects a more or less parallel arrangement of the biotite and a tendency for the biotite to occur in fairly well-defined bands associated largely with quartz (Fig. 3, *C*). Some seams of quartz also tend to lie along and accentuate the banding. However, the andesine grains are not aligned nor do they approach any kind of parallel arrangement.

Much of the andesine shows about the same features as it does in the biotite-quartz diorite. Where it is most abundant in the rock its grains are euhedral to subhedral, but where it is not so abundant its grains are notably irregular. Characteristically, most of them hold rounded, remnant inclusions of quartz and biotite, inherited from more or less micaceous quartzite. In some places remnants of the micaceous quartzite of appreciable size are held in the gneiss (Fig. 3, *C*). These inclusions show that the oriented grains of biotite have replaced the quartzite. Widely scattered crystals of andesine also may be embedded in these quartzitic inclusions and may extend into and replace the quartz and biotite more or less idiomorphically, holding each of them as rounded or irregular remnant inclusions. In general the biotite is concentrated along the more quartzose bands of the gneiss, rather than along the feldspathic; and, where the andesine invades the quartzitic bands, the unreplaced parts of the biotite grains tend to mantle the andesine crystals. However, not all the biotite in the gneiss is that associated with and inherited from the quartzite, but there are also grains of younger biotite which penetrate into and through the andesine. The quantity of the younger biotite is in general very much less than that which was originally present in the more quartzose but biotite-rich bands. Because the biotite is lost through replacement, its content appears to decrease proportionately as the content of andesine increases.

In addition to the remnant, rounded, and irregular quartz grains locked in the andesine crystals, there are also variable but generally large amounts of quartz in lobate grains and in irregular veinlets (Fig. 3, *D*) which penetrate the andesine and biotite and which in places engulf remnants of them and the early quartz. This quartz has precisely the same relations as the late quartz in the biotite-quartz diorite. Some of the quartz forms comparatively long seams

commonly parallel to the foliation of the rock. Such seams may pass through grains of biotite and andesine as well as remnants of the early quartz. The amount of quartz added in this manner, however, is variable and may locally exceed or be exceeded by quartz of earlier age.

The microcline, which occurs in some of the rock, is in relatively small grains. These show a tendency to penetrate into and replace the andesine and to hold remnants of the andesine and early quartz as irregular to rounded inclusions. In places where the microcline is in contact with the andesine, the latter contains myrmekite-like blebs of quartz. The microcline is to some extent penetrated by lobate quartz.

The only accessories of notable abundance are the apatite, zircon, magnetite, and garnet. These tend to be concentrated in lines parallel to the foliation and commonly with that biotite which appears to be younger than the andesine. The accessories show some preference for grain contacts of other minerals. Here and there muscovite replaces andesine and forms small patches in it. Epidote, sphene, and allanite are observed but rarely. Zoisite in large, rounded, and rectangular grains are embedded in andesine and biotite.

AUGEN GNEISS

The augen gneiss is invariably associated with the quartz-biotite-diorite gneiss and differs from it only in containing microcline porphyroblasts which are as much as $\frac{1}{2}$ inch thick and $1\frac{1}{2}$ inches long. As the porphyroblasts are more or less ovoid or form lenslike grains oriented with the long dimension parallel to the foliation, the gneiss has a lenticular rather than a banded texture. However, a few of the porphyroblasts tend to have the shapes of euhedral crystals and give a more spotted appearance to the rock. The matrix between the porphyroblasts does not differ materially from the rock composing the quartz-biotite-diorite gneiss. The biotite bands tend to lie between and to wrap around the porphyroblasts, but some of the biotite appears as shadows in and through them. The porphyroblasts seem to have sharp borders in the hand specimen; but actually they are minutely irregular and penetrate deeply and unevenly into the bordering biotite, plagioclase, and quartz grains. The list of

minerals of the augen gneiss is the same as that of the other gneiss; the only difference is the larger amount of microcline and the tendency of some of it to form large "augen."

Early grains of quartz and biotite, inherited apparently from the quartzite and penetrated by and included in the andesine, appear in the augen gneiss as they do in the quartz-biotite-diorite gneiss. Also, the microcline porphyroblasts contain an abundance of shadow-like inclusions and actual remnants of andesine, early quartz, and biotite which were not entirely replaced by them (Fig. 4, *A*). Although some of the porphyroblasts have borders of what appears to be more or less granulated material, there is no evidence of crushing or that the porphyroblasts are phenocrysts oriented by granulation. What appears to suggest granulation has been produced not by crushing but by thorough impregnation and interlacing of the andesine, biotite, and early quartz by continuous and discontinuous veinlets and masses of microcline (Fig. 4, *B*). Out of reach of the microcline porphyroblasts, the apparent granulation disappears (Fig. 4, *C*). The granulation seems to be a pseudoclastic effect and the result of replacement of the bordering minerals by microcline. It appears to be an example of the pseudoclastic process of replacement described by G. H. Anderson in rocks which contain late microcline.[8] The matrix in general contains much microcline; but there are places where it is absent. Here the rock or matrix is essentially a micaceous and somewhat feldspathic quartzite (Fig. 4, *C*).

Biotite younger as well as older than the andesine is present in the rock. Lobate quartz grains and penetrating veinlets are also present, but the younger quartz is not nearly so conspicuous as in the quartz-biotite-diorite gneiss. Much of the late quartz is in the matrix and does not penetrate the porphyroblasts to any appreciable extent. Muscovite in small grains replaces some of the andesine; less commonly, the microcline. Accessories show spatial distribution and relation to grain boundaries as in the other rocks. The only other feature worthy of comment is an abundance of myrmekite-like quartz in andesine grains invaded by microcline. These inter-

[8] "Pseudo-cataclastic Textures of Replacement Origin in Igneous Rocks," *Amer. Mineralogist*, Vol. XIX (1934), pp. 185–93.

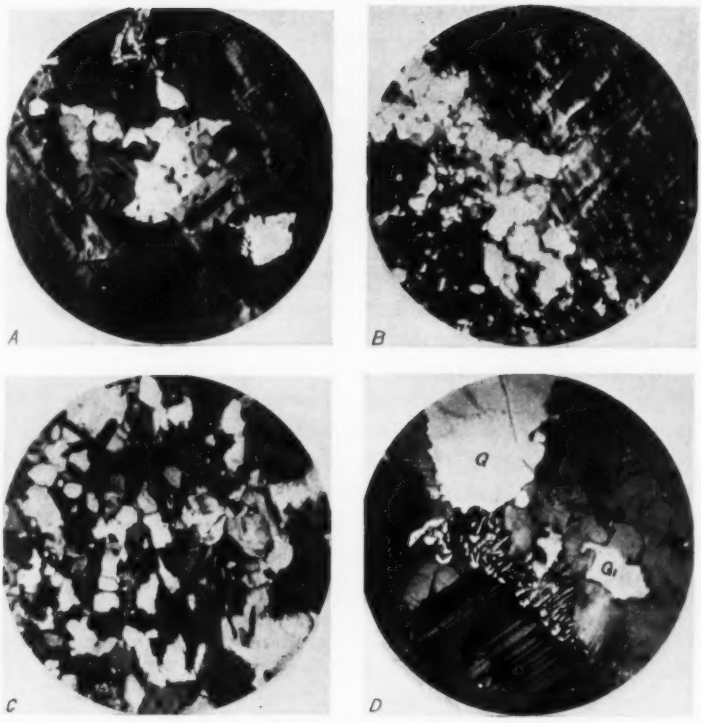

FIG. 4.—*A*, Photomicrograph of the augen gneiss showing a microcline porphyroblast with numerous remnants of plagioclase, quartz, and biotite inherited from the earlier gneiss. These inclusions indicate that the porphyroblasts have been formed late in the gneiss by replacement. Mag. 21×. *B*, Photomicrograph of the augen gneiss showing the minutely irregular borders of the microcline porphyroblasts and the manner in which the microcline in continuous and discontinuous veinlets and masses penetrates and interlaces with the andesine, biotite, and early quartz of the adjacent matrix. The apparent granulation is a pseudoclastic effect produced by replacement of the bordering minerals by the microcline. Mag. 21×. *C*, Photomicrograph of the augen gneiss showing the character of the matrix between microcline porphyroblasts. Biotite is oriented and is associated with quartz as in the regular quartz-biotite-diorite gneiss. Crystals of andesine and grains of microcline also occur in the matrix. Mag. 21×. D, Photomicrograph of pegmatite showing andesine, quartz (Q_i), and biotite (*B*), inherited from the biotite-quartz diorite wall rock, corroded and partly replaced by microcline (*M*), and the microcline in turn invaded by younger lobate quartz (*Q*). Myrmekitic-like quartz blebs occur in the andesine which has been corroded by the microcline. Such quartz-plagioclase intergrowths are abundantly developed in the granodiorite and gneiss or wherever microcline has come in contact with andesine. Mag. 21×.

growths are apparently a more or less characteristic feature of the augen gneiss.

The method by which the microcline porphyroblasts have become enclosed in biotite is not altogether clear. The microscope confirms evidence gathered from the hand specimen that the biotite bands tend to bend around the "augen" but gives no clue as to why the biotite, which is older than the microcline, should occupy that position.

GRANODIORITE

Occurring as a broad but irregular band between the gneissic rocks and the biotite-quartz diorite, the granodiorite has some of the features of each; and it may greatly resemble or be practically indistinguishable in the hand specimen from the bordering rocks. That type which is practically indistinguishable from the biotite-quartz diorite differs from it only in containing appreciable amounts of microcline. That type which has some resemblance to the quartz-biotite-diorite gneiss differs in having notable amounts of microcline but only a faint or generally obscure banding.

Estimates of the usual proportions of essential minerals are andesine, 40–50 per cent; microcline, 15–20 per cent; quartz, 25–35 per cent; and biotite, 5–10 per cent. Other minerals are accessory zircon, apatite, magnetite, muscovite, and garnet; and such other minerals as zoisite, sericitic mica, chlorite, and calcite.

Whether the rock resembles the biotite-quartz diorite or the quartz-biotite-diorite gneiss, the mineral relations are essentially the same. The andesine generally has phantom or real inclusions of quartz, less commonly of biotite; and in places, also, residua of an older, more calcic, plagioclase. The andesine with the more calcic plagioclase inclusion has exactly the same characteristics as the andesine in the biotite-quartz diorite. The andesine which contains inclusions of quartz and biotite has the same characteristics as that in the gneiss. As in the other rocks, the andesine is penetrated by biotite of younger age and is extensively cut and embayed by grains of microcline. Its grains also are engulfed in the microcline as scattered but oriented island inclusions. The microcline also may hold rounded grains of quartz and, usually, remnants of biotite. Some microcline grains may hold a host of the earlier minerals in the same

ways as do the microcline porphyroblasts in the augen gneiss (Fig. 4, *A*). Andesine in contact with and corroded by the microcline not uncommonly contains blebs of myrmekitic-like quartz. The andesine and biotite are extensively embayed and cut by lobes and veinlets of quartz; the microcline, not so conspicuously so. Permeation of the granodiorite by late quartz is as widespread and evident as it is in the other rocks of the area.

Muscovite is notably abundant in the granodiorite. Some of it occurs in reticulating veinlets along and across the cleavage in the andesine. It appears more rarely in the microcline. Zircon, apatite, garnet, and magnetite tend to be concentrated along contacts between other minerals and along cleavages and fractures in them. Zoisite forms rounded grains in biotite and andesine. Sericitic mica, chlorite, and calcite are present and have the same relations as in other rocks already described. Except for the presence of rather abundant microcline, the granodiorite differs very little from either the biotite-quartz diorite or the quartz-biotite-diorite gneiss in minerals and mineral relationships.

PEGMATITES

The pegmatites are moderately to fairly coarse-grained rocks with distinctly granitoid textures. With few exceptions the most conspicuous mineral is black biotite, which occurs through the rock in scattered crystals from $\frac{1}{4}$ inch to 3 inches or more in diameter. Some of the feldspar crystals are as much as 2 inches long, but the other mineral grains are usually somewhat smaller. Many of the pegmatites are highly feldspathic; others, highly siliceous. Their composition varies from place to place, but, in general, is between biotite-granite pegmatite and biotite-granodiorite pegmatite. Borders of the pegmatites are not sharply defined but shade more or less gradually into the adjacent rock.

Microcline is the most plentiful mineral in all the course-grained pegmatites; in those with finer grain, andesine may exceed it. Quartz also may exceed either but is ordinarily subordinate. Minerals other than the essential biotite, microcline, andesine, and quartz include zircon, apatite, magnetite, and muscovite, in some places albite, and locally garnet. In one pegmatite, garnet is the most abundant mineral.

The andesine and a part of the quartz are penetrated by or are engulfed as rounded and irregular remnants in the microcline. None of the microcline crystals lacks the inclusions (Fig. 4, *D*), which become increasingly numerous near the margins of the pegmatites and eventually unite with grains of the biotite-quartz diorite and granodiorite wall rock. The penetrating and veining habit of the microcline is best shown in the finer-grained varieties of granodioritic composition. The engulfing relations are typical of the coarse-grained rock in which the microcline grains are large enough to hold numerous andesine and quartz remnants or their ghostlike outlines. As the included little-replaced andesine and quartz grains correspond in size with those of the invaded dioritic or granodioritic rock, and as they merge with them in the transitional or gradational zone, they have obviously been inherited from the wall rock and are but unreplaced residua of the invaded biotite-quartz diorite and granodiorite. Much myrmekite-like quartz appears wherever the microcline comes in contact with the inherited andesine (Fig. 4, *D*).

Some of the smaller biotite grains in the finer-grained pegmatites have apparently, like the andesine and some of the quartz, been inherited from the wall rock; but the large crystals extend through and replace the andesine and quartz and even project into the bordering biotite-quartz diorite and granodiorite. This later biotite has not been penetrated by microcline, but it, as well as the microcline and andesine (with associated quartz), have been embayed and penetrated by grains of lobate quartz (Fig. 4, *D*) and cut by irregular veinlets of the quartz. Most of the quartz in the pegmatites is this late variety and replaces earlier minerals extensively. It has been introduced in variable amounts in the different pegmatites and also in different parts of the same pegmatite.

Some of the microcline is perthitic and is cut by irregular veinlets of albite in part parallel to 100. Some larger crystals of albite also have been observed in some of the pegmatites, and these extend into and replace not only the microcline but also some of the quartz.

Muscovite is visible in but few places as small grains in and replacing andesine and microcline. Magnetite, zircon, and apatite are present in scant amounts in microcline and biotite, and most of the grains appear to have been introduced more or less at the same time as the late quartz. Some grains were probably inherited from the

invaded rock. In the garnet pegmatite, which is about 4 feet thick, the pale red garnet has replaced most of the late quartz and most of the feldspar that might originally have been present.

SILEXITE

The dikes and lenticular pods of silexite are composed of milky to glassy quartz accompanied by a few widely scattered grains of muscovite and biotite. The bodies are essentially monomineralic, as the other minerals ordinarily make up less than 1 per cent of the rock. The silexite is moderately coarse grained and in the hand specimen is somewhat difficult to distinguish from the quartzite. In some places the silexite shows transitions into pegmatite and may have as much as 30 per cent microcline present as sharp, almost rectangular, crystals an inch or more long. A few of the silexite bodies have cavities lined with glassy quartz crystals.

QUARTZ GABBRO

The quartz gabbro is medium to fine grained, grayish to pale green, and is weakly to prominently foliated. It commonly contains evenly spaced seams of quartz along the foliation planes. The most abundant minerals are quartz and augite; but bytownite, though more or less subordinate to them, is an essential mineral. Microcline is present in places, and there are exceptionally large amounts of zircon. Some apatite, magnetite, and sphene, as well as some uralitic hornblende as an alteration product of augite, are also present. The proportions of different minerals vary; but in one of the most typical specimens there are about 40 per cent quartz, 30–35 per cent augite, 15–20 per cent bytownite, and about 5 per cent microcline.

The augite and bytownite occur as anhedra which have been penetrated extensively by quartz and locally by zircon grains and crystals, some of them of rather large size. The augite and bytownite also have been embayed, engulfed, and replaced by microcline grains. Much of the quartz is in fairly large grains and veinlets parallel to the foliation and embaying the minerals unevenly on either side; but some is in smaller grains which lie between but embay the crystals of augite and bytownite.

Because the quartz gabbro is extensively silicified and has been penetrated by considerable amounts of microcline and zircon, it is

regarded as older than the batholith. Probably it represents engulfed, only partly "granitized" inclusions in the biotite-quartz diorite and granodiorite.

PETROLOGY

SIGNIFICANCE OF MINERAL RELATIONS

Petrographic study of the rocks on Paradise Ridge has shown that there is a marked similarity in the kinds of minerals and in their relationships, not only in the igneous rocks but also in the invaded country rock. This similarity is apparent especially in the manner and order in which the minerals formed in the igneous and metamorphic rocks. For example, much of the andesine in the biotite-quartz diorite contains inclusions of quartz and biotite or of quartz, biotite, and a more calcic andesine. Much of the andesine in the gneiss contains remnant grains of quartz and biotite inherited from quartzite. In either rock the andesine is somewhat penetrated and replaced by biotite. Wherever microcline appears it penetrates, engulfs, and retains whole masses of andesine, earlier quartz, and biotite as remnants or as shadow-like inclusions. On the other hand, much of the quartz, which occurs in the form of lobate grains and irregular veinlets, embays and cuts the biotite and feldspars. It penetrates the quartzite in a similar manner. Myrmekitic-like quartz appears here and there wherever the microcline comes in contact with grains of andesine, a phenomenon found in the gneissic rocks as well as in the granodiorite and pegmatites. The various accessory minerals, such as zircon, apatite, garnet, and magnetite, mainly occur near or along fractures, grain borders, and cleavages in the biotite and feldspars, and commonly are found near the lobate grains and veinlets of quartz. Muscovite shows a somewhat similar distribution along grain contacts and cleavages but replaces the feldspars. From these relations it is apparent that each rock has been affected by mineral replacements in much the same way and order as any other rock. The relations suggest that for the most part both the invading and the invaded rocks have been greatly modified by replacements and that some have been formed altogether by replacement of pre-existing rocks.

ORIGIN OF ROCK TYPES

Hornblende-biotite diorite.—The hornblende-biotite diorite is apparently the only rock which has not been formed by the replace-

ment of some other rock. Its early hornblende, biotite, and rather calcic andesine contain no remnants of other minerals but appear to have crystallized from the rather quickly chilled magma along the margin of the batholith. Nevertheless, the hornblende-biotite diorite has been changed greatly in places by the addition of quartz, accessory minerals, and considerable amounts of epidote. Inclusions of hornblende-biotite diorite in the younger biotite-quartz diorite and granodiorite show, by the loss of hornblende and by the gain of a less calcic andesine through replacement of the original more calcic one, various stages of transition into the biotite-quartz diorite. By an additional gain of microcline the rock shows stages of transition into granodiorite. The completed transformation is into biotite-quartz diorite and locally into granodiorite.

Biotite-quartz diorite.—Although the biotite-quartz diorite is regarded as the characteristic consolidated rock of the marginal zone of the batholith,[9] the andesine in a considerable part of the rock on Paradise Ridge invariably contains distinct remnants of an earlier hornblende-biotite-quartz-plagioclase rock or of a quartz-biotite rock. The remnants of the former are identical with the hornblende-biotite diorite of the chilled border shell; the latter, with the quartzite inclusions of the quartz-biotite gneiss. The evidence seems clear that the biotite-quartz diorite has been formed at least in part by replacement of the earlier and more calcic border of the batholith and by replacement of the somewhat micaceous quartzite. The rock also has been modified considerably by the addition of biotite and especially large amounts of quartz, which have made it exceptionably quartz-rich. Various accessories also were added, and, locally, a little muscovite.

Quartzite.—The quartzite was formed apparently from a relatively pure sandstone, the original grain borders of which were obliterated entirely through recrystallization and the resulting interlocking anhedra further modified by the superposition of penetrating seams and lobate grains of quartz. Through a part of the quartzite other minerals were added in advance of the quartz. These earlier additions included biotite first and then andesine, which by their addition changed some of the quartzite to biotite diorite and, where the

[9] A. L. Anderson, "The Geology and Mineral Resources of the Region about Orofino, Idaho," *loc. cit.*

biotite grains were aligned in parallel to subparallel position, presumably under the influence of differential stress, to gneiss. The quartz, which was added later, permeated throughout the body of the quartzite and caused it and its "granitized" parts to be notably silicified.

Quartz-biotite-diorite gneiss.—The quartz-biotite-diorite gneiss is but a facies of the "granitized" quartzite. Its andesine contains inclusions of early quartz and biotite, just as the andesine of some of the biotite-quartz diorite contains inclusions of them and of the minerals of the earlier chilled diorite. In addition, there are remnants of numerous bands of the biotite-bearing quartzite impregnated to greater or lesser degree by the andesine. Except for this, its history is not much different from that of the biotite-quartz diorite; some biotite, potash feldspar, muscovite, and numerous accessory minerals were added, and later still, large amounts of quartz. Although the later replacing minerals apparently were not formed as aligned stress minerals, their random orientation did not destroy the foliation of the rock which had evolved under stress at the time the early biotite was deposited. However, in places the rock is distinguished with some difficulty from the biotite-quartz diorite.

Augen gneiss.—The augen gneiss is quartz-biotite-diorite gneiss to which microcline has been added in abundance. Because the microcline tended to form as large grains more or less parallel to the foliation, the quartz-biotite-diorite gneiss was made more alkalic and changed to augen gneiss. Otherwise the augen gneiss has similar relicts of quartzite impregnated by biotite and andesine as are found in the quartz-biotite-diorite gneiss. There are also grains of younger biotite, quartz, and accessories.

Granodiorite.—The granodiorite which merges with the quartz-biotite-diorite gneiss, on the one hand, and the biotite-quartz diorite, on the other, differs from each only in containing appreciable amounts of microcline. The granodiorite shows little or no evidence of banding. Most of it has been formed by the addition of microcline to the biotite-quartz diorite; some of it, by the addition of microcline to the quartz-biotite-diorite gneiss. Like the other rocks the granodiorite was silicified rather extensively and had small but variable amounts of accessories and muscovite added to it.

Pegmatite and silexite.—The pegmatites have been formed by re-

placement of the biotite-quartz diorite and granodiorite. The biotite, microcline, quartz, and, locally, albite and garnet, which form the pegmatites, invaded and replaced the minerals of the bordering country rock. The process of pegmatitization was not unlike the processes which were active during the earlier endomorphism of the granite rock, but the solutions apparently acted after consolidation and general endomorphism had gone to considerable depth and the solutions had become fairly concentrated. The silexites appear to represent rocks in which late silicification was complete or where the late quartz was deposited in fissures.

Quartz gabbro.—The quartz gabbro is a rock which has been extensively silicified and to some extent replaced by microcline and particularly zircon. No gabbro was observed in the quartzite; but its occurrence in the quartz diorite and granodiorite could well be interpreted as inclusions which had been metamorphosed more or less by siliceous emanations from the invading magma.

ROLE OF HYDROTHERMAL REPLACEMENT

The changes which have taken place in the rocks and which to a marked degree have controlled their composition have resulted from replacement of pre-existing mineral grains by an assemblage which is regarded as hydrothermal. Such replacements acting through the agency of high-temperature hydrothermal solutions could have been associated only with and derived from the batholithic magma. As the magma is regarded commonly as a solution—fluid mainly through heat—its substance cannot enter readily into minute openings by which a replacing solution permeates a rock and, therefore, it cannot cause widespread replacement. As hydrothermal solutions derived from and gradational into the parent magma owe their fluidity chiefly to the presence of aqueous constituents, they can penetrate readily and can cause extensive changes by replacements. For this reason, it is believed that hydrothermal emanations associated with the batholithic magma permeated the rocks above and, by the addition of material contained in solution, caused widespread changes, including all the observed replacements and exomorphic and endomorphic phenomena.

As the fluids penetrated the quartzite along zones of special permeability, determined by shearing under differential stress, the quartzite became impregnated with biotitic and feldspathic constitu-

ents and was changed to quartz-biotite-diorite gneiss, in part to biotite-quartz diorite, and, where the solutions were enriched later in potash, to augen gneiss. In the same way much rock of the early chilled border of the batholith, penetrated by the emanations, was changed largely to biotite-quartz diorite; and the rock alongside the augen gneiss was changed to granodiorite. Then, as the emanations continued to permeate upward, the rocks, regardless of their original character, were extensively silicified and had various accessory minerals added to them. As hydrothermal solutions from greater depths ascended along fractures and fissures in the already endomorphosed biotite-quartz diorite and granodiorite and soaked into the rock alongside, pegmatites and to some extent silexites were formed by replacement.

GENERAL RECOGNITION OF HYDROTHERMAL REPLACEMENTS IN EXOMORPHISM AND ENDOMORPHISM

That hydrothermal replacements may be a factor in the formation of igneous and metamorphic rock types has been recognized by many observers. Freeman has described recently replacement shells around batholiths in the Waswanipi district in northwestern Quebec in which hydrothermal emanations from a solidifying batholith permeated and altered the invaded rocks to gneiss.[10] Fenner at a rather early date pointed out the influence of magmatic emanations in the formation of certain gneisses in the Highlands of New Jersey.[11] Somewhat similar phenomena have been described by Lacroix,[12] Adams and Barlow,[13] Quirke and Collins,[14] Stark,[15] and Barth.[16]

[10] B. C. Freeman, "Replacement Shells around Batholiths in the Waswanipi District Northwestern Quebec," *Jour. Geol.*, Vol. XLVI (1938), pp. 681–99.

[11] C. N. Fenner, "A Mode of Formation of Certain Gneisses in the Highlands of New Jersey," *Jour. Geol.*, Vol. XXII (1914), pp. 697–702.

[12] A. Lacroix, "Le Granit des Pyrénées et ses phenomènes de contact," *Carte géol. de France*, 1900, p. 26.

[13] F. D. Adams and A. E. Barlow, "The Geology of the Haliburton and Bancroft Areas, Ontario," *Geol. Surv. Canada Mem. 6* (1910), pp. 81–83, 179–80.

[14] T. T. Quirke and W. H. Collins, "The Disappearance of the Huronian," *Geol. Surv. Canada Mem. 160* (1930).

[15] J. T. Stark, "Migmatites of the Sawatch Range, Colorado," *Jour. Geol.*, Vol. XLIII (1935), pp. 1–26.

[16] T. F. W. Barth, "Structural and Petrologic Studies in Dutchess County, New York, Part II: Petrology and Metamorphism of the Paleozoic Rocks," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 775–850.

Clearly defined contact phenomena manifested by the impregnation of "igneous" material into quartzite in such a manner as to make it resemble granitic rock has been described by the senior author.[17] Goodspeed has described recently an occurrence of granodiorite which has resulted from the metamorphism of hornfels,[18] and Bastin has shown that the dikelike aplites with the Nipissing diabase were formed by hydrothermal replacements.[19]

That the escaping emanations may bring about endomorphism of the invading rock itself has been observed in many places. Gillson described such changes in the granodiorites of the Pend Oreille district of Idaho[20] and in the Pioche district of Nevada,[21] G. H. Anderson has described such phenomena in the Northern Inyo Range of California-Nevada.[22] Currier also has described extensive hydrothermal replacements in a part of the Idaho batholith in the Yellow Pine district, Idaho.[23] Extensive changes in the consolidated rock of a larger part of the Idaho batholith by post-consolidation emanations are to be discussed in some detail in a paper now in preparation.[24] An interesting example wherein quartz diorite has been changed to albite granite by hydrothermal replacement is described in detail by Gilluly.[25]

SUMMARY AND CONCLUSIONS

Because there is such a close genetic relationship among the rocks of Paradise Ridge, the members can be explained only as the prod-

[17] A. L. Anderson, "Contact Phenomena Associated with the Cassia Batholith, Idaho," *loc. cit.*

[18] G. E. Goodspeed, "Small Granodiorite Blocks Formed by Additive Metamorphism," *Jour. Geol.*, Vol. XLV (1937), pp. 741–62.

[19] E. S. Bastin, "'Aplites' of Hydrothermal Origin Associated with Canadian Silver-Cobalt Ores," *Econ. Geol.*, Vol. XXX (1935), pp. 715–34.

[20] J. L. Gillson, "Granodiorites of the Pend Oreille District of Northern Idaho," *Jour. Geol.*, Vol. XXXV (1927), pp. 1–27.

[21] "Petrography of the Pioche District, Lincoln County, Nevada," *U.S. Geol. Surv. Prof. Paper 158-D* (1929), pp. 77–86.

[22] "Granitization, Albitization and Related Phenomena in the Northern Inyo Range of California-Nevada," *Geol. Soc. Amer. Bull.*, Vol. XLVIII (1937), pp. 1–74.

[23] L. W. Currier, "A Preliminary Report on the Geology and Ore Deposits of the Eastern Part of the Yellow Pine District, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 43* (1935), pp. 7–11.

[24] A. L. Anderson, "Endomorphism of the Idaho Batholith." In preparation.

[25] James Gilluly, "Replacement Origin of the Albite Granite near Sparta, Oregon," *U.S. Geol. Surv. Prof. Paper 175-C* (1933), pp. 65–81.

ucts of a single cycle of igneous activity beginning with the intrusion of magma into essentially pure quartzite. The intrusion of magma was accompanied at first by rapid chilling, which gave rise to a shell of dark, fine-grained hornblende-biotite diorite. As a result of the early chilling and rapid crystallization, it is likely that little or no emanations escaped from the magma during the early stages of its consolidation, but some time later, after the solidified border shell had been shattered, it and the fractured quartzite above were permeated by hydrothermal solutions associated with and escaping from the solidifying magma at somewhat greater depth.

As the hydrothermal solutions soaked through the rock above, they brought about a series of transformations. The solutions caused some of the quartzite and most of the consolidated hornblende-biotite diorite to be replaced by biotite and andesine, changing the one into a biotite-diorite gneiss and in part to biotite-diorite and the other into a less calcic diorite. As the emanations continued to stream upward, they induced further changes in the rock by causing the addition of more biotite; and, because the solutions ascending the main zone of structural weakness along and through the gneiss had become enriched in potash, they also added locally much microcline. This microcline changed much of the biotite-diorite gneiss to augen gneiss, distinguished by many conspicuous microcline porphyroblasts. The potassic emanations also permeated the bordering biotite diorite and caused some of it to be changed to granodiorite. Then, as the solutions became more notably siliceous and fluid, perhaps in part from silica taken into solution as a result of earlier replacements, they permeated all the rocks and deposited everywhere by replacement exceptionally large amounts of quartz. The enrichment in silica changed the biotite diorite to a quartz-rich biotite-quartz diorite, some of the remnants of hornblende-biotite diorite, to hornblende-biotite-quartz diorite, and the probable inclusions of gabbro to quartz gabbro. The amount of quartz in the biotite-diorite gneiss and in the granodiorite likewise was increased, and the quartzite was extensively silicified. At about the same time minor amounts of muscovite were formed in some of the rocks. Also, generally small amounts of sphene and considerable amounts of zircon, apatite, magnetite, and garnet were added by replacement along fractures, cleavages, and grain borders of the earlier minerals. The

late solutions also caused formation of some allanite, zoisite, epidote, chlorite, sericite, and calcite.

These exomorphic and endomorphic modifications appear to reflect progressive changes in the composition of the pervasive emanations, changes of composition which accord more or less with declining temperatures.

After the rocks had ceased to be affected by these emanations, deeper-seated fracturing apparently tapped hydrothermal solutions from lower levels and guided them upward into the joints and fissures in the solid, endomorphosed rock above. As the solutions spread into the rock bordering the fractures, they added material and caused the rock to be replaced by variable but generally coarse-grained aggregates of biotite, microcline, and quartz, thereby changing the rock to pegmatite. In places the later solutions became somewhat sodic, and variable but small amounts of albite also were added by replacing the feldspar and quartz. In one pegmatite so much late garnet was added that much of the earlier feldspar and quartz were eliminated. Locally a surplus of silica in the late fluids was deposited as quartz, giving rise to bodies of silexite, which filled fractures or replaced the bordering rocks.

Thus the rocks on Paradise Ridge have had a vastly different history of development from the rocks of similar kind in the Bald Ridge Butte area 9 miles away, where normal conditions of cooling and crystallization have been postulated.[26] It is altogether likely that the exomorphic processes noted in the gneissic rocks on Paradise Ridge also may have played a similar part in the formation of crystalline schists, especially the quartz-biotite-plagioclase schists, in extensive metamorphic regions. Freeman's observations and deductions in the Waswanipi district in Quebec tend to show that this is true.[27] The Paradise Ridge study shows, too, that augen gneisses are not necessarily cataclastic gneisses but that porphyroblastic and pseudo-cataclastic effects may be produced by replacement.

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[26] Hoffman, "The Geology of Bald Ridge Butte, Washington," *op. cit.*, pp. 647–50.

[27] "Replacement Shells around Batholiths in the Waswanipi District Northwestern Quebec," *loc. cit.*

STRUCTURE OF THE PRYOR MOUNTAINS MONTANA

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ABSTRACT

The asymmetric "trap-door" type of uplifts which are significant and characteristic features of the structure of the Central Rockies are developed in a genetic sequence in the Pryor Mountain uplift. Excellent surface expression with topographic relief of canyon type combine to make the Pryor uplift a favorable place to determine the three-dimensional form and mode of origin of this type of uplift.

The writer's study of the Pryor uplift indicates that these asymmetric anticlines are due to tangential compression acting on curved fault planes in the basement complex. Elevation and rotation of the segments preceded thrusting over the steep limb of the fold. This structural phenomenon is similar to that described by Willis as "ramp" structure. The intervening basin area between the Pryor Mountains and the northern Big Horn uplift is compared to a "ramp "valley.

The structural interpretation offered in this paper is an attempt to explain the overthrusting associated with asymmetric anticlines of the Central Rocky Mountain region. This overthrusting has been avoided in previous explanations of similar anticlines. An explanation based on normal faults activated by gravity does not seem applicable.

INTRODUCTION

The Pryor Mountain uplift of southern Montana forms a typical unit within a regionally extensive pattern of the Central Rocky Mountains. The study of the Pryor Mountains was undertaken as a part of the research program conducted in the Beartooth-Bighorn-Yellowstone region in order that data concerning their form and makeup could be applied to the solution of major problems of Rocky Mountain orogeny.

The emphasis of this investigation was placed upon the structure of the mountain mass. Because the stratigraphy of the region is relatively well known, the writer's attention was centered chiefly upon determining the form and space relationships of the crustal blocks comprising the uplift. Only such studies of the lithology and fossil content of the formations were made as were essential to the execution of the geologic mapping and structural studies.

The area mapped, which includes approximately 450 square miles, lies principally in south-central Montana (Fig. 1) but extends a short distance into Wyoming. The mountain mass forms part of the northeastern margin of the Bighorn Basin and is contiguous to the

northern portion of the Bighorn Range. The map area, bounded by 44°55′ and 45°25′ north latitude and 108°10′ and 108°40′ west longitude, lies in three counties—Carbon and Bighorn, Montana, and Bighorn, Wyoming. There are no towns or main highways in the Pryor uplift.

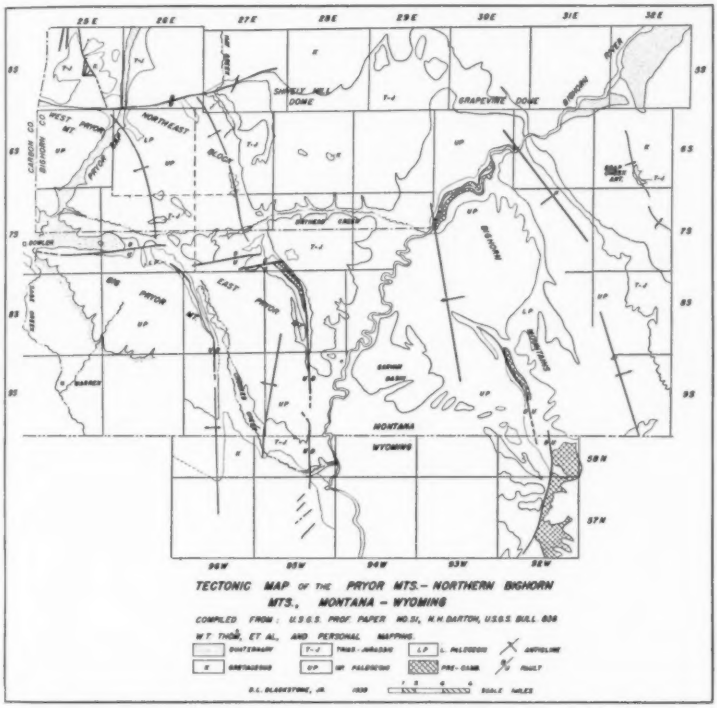

FIG. 1.—Tectonic sketch map of the Pryor Mountains and adjacent areas

Topographically, the area may be best described as a series of moderately dissected block mountains drained by the Bighorn and Clark Fork rivers and their tributaries. The gentle dip slopes to the west are eroded into narrow box canyons, while the steeper eastern slopes of the blocks are abrupt scarps, though relatively less dissected. The mountain mass is separated from the northern portion

of the Bighorn Range by a structural trough, through which the Bighorn River flows in deeply incised meanders. The extreme northwest portion of the Pryor Mountains is separated from the remainder of the mass by the wind gap known as Pryor Gap—the abandoned channel of an earlier stage of the Shoshone River, as described by Mackin.[1]

The field work embodied in this paper was carried out during portions of the field seasons of 1932–35, inclusive. The mapping was accomplished principally by means of plane-table and telescopic alidade or upon a United States Forest Service base map covering the portion of the area within the Pryor division of the Custer National Forest.

The Pryor Mountains were first referred to by Lewis and Clark[2] on the return from their expedition to the Pacific, and the mountains have derived their name from Pryor Creek, so named after Sergeant Pryor of the Lewis and Clark expedition. Previous geologic work in the area has been limited to the margins of the uplift.[3] In particular, the work by R. S. Knappen and G. F. Moulton and by W. T. Thom, Jr., G. M. Hall, C. H. Wegemann, and G. F. Moulton[4] has been of great aid in recognizing the stratigraphic units and marginal features of the uplift. Thom[5] and Dobbin also prepared a generalized structure-contour map of the Northern Plains upon which the Pryor uplift was depicted. This map was later modified and augmented by C. E. Dobbin and C. E. Erdmann.[6]

[1] J. Hoover Mackin, "The Erosional History of the Big Horn Basin, Wyoming," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 813–94.

[2] R. G. Thwaites, *Original Journals of the Lewis and Clark Expedition, 1804–1806*, Vol. V (July 24, 1806), p. 290.

[3] G. H. Eldredge, "A Geological Reconnaissance of Northwest Wyoming," *U.S. Geol. Surv. Bull. 119* (1894), p. 2; C. A. Fisher, "Geology and Water Resources of the Bighorn Basin, Wyoming," *U.S. Geol. Surv. Prof. Paper 53* (1906); N. H. Darton, "Geology of the Bighorn Mountains," *U.S. Geol. Surv. Prof. Paper 51* (1906).

[4] "Geology and Mineral Resources of Parts of Carbon, Big Horn, Yellowstone, and Stillwater Counties, Montana," *U.S. Geol. Surv. Bull. 822-A* (1928); "Geology of Big Horn County and the Crow Indian Reservation, Montana," *U.S. Geol. Surv. Bull. 856* (1935).

[5] W. T. Thom, Jr., "The Relation of Deep-Seated Faults to the Surface Structural Features of Central Montana," *Bull. Amer. Assoc. Pet. Geol.*, Vol. VII (1923).

[6] "Structure Contour Map of the Montana Plains," *U.S. Geol. Surv.* (1935).

TABLE 1

FORMATIONS OUTCROPPING IN THE PRYOR MOUNTAINS, MONTANA

		FEET
Quaternary		
Recent.....	Alluvial fans, stream deposits	
Pleistocene.....	(?) Locally derived limestone gravels occupying pediments. Small amounts of gravels, containing pebbles of volcanic material not locally derived	
	Unconformity	
Cretaceous		
Upper.....	Mowry shale	
	Thinly bedded, siliceous shale, weathering silvery, contains numerous fish scales	250
	Thermopolis shale	
	Black shales; black sandy shales; black shales with bentonite; and rusty sandy shale, or sandstone, strongly ripple marked	650
	Cloverly formation	
	Greybull sandstone member—sandy shale or sandstone	20
	Variable clays of red, purple, and pink color. Bone scrap and gastroliths (?) common	130
	Pryor conglomerate member. Conglomerate or coarse-grained sandstone, changing to latter from north to south. Black chert pebbles or fragments in a quartzitic matrix	50
Total.....		200
	Unconformity	
Jurassic		
Upper.....	Morrison formation	
	Gray, very gritty shale, weathering into soft slope. Lenses of red shale 20′–25′ thick at base	170
	Yellow buff sandstone, medium grained	70
Total.....		240
	Unconformity	
	Sundance formation	
	Olive green sandstone, with *Gryphea*, and *Belemnites*	40
	Gray, soft shales, with minor sandstones and sandy shales	115
	Gray and yellow shales, thinly bedded sandstones, bentonitic shale at top	120
	Gray, fossiliferous limestone	3
Total.....		278
	Unconformity	
Triassic.....	Chugwater formation	
	Drab maroon shale, with gypsum streaks	100
	Gray, dense, massive limestones, with maroon shale partings, and limy streaks	14
	Red sandy shale, and sandstone	110
	Granular gypsum (traced to north this grades into a three-foot thinly bedded limestone)	60
	Red sandy shales, and sandstone	330
Total.....		614
	Unconformity	

TABLE 1—*Continued*

		FEET
Permian (?).....	Embar (?)	
	Absent on west side of uplift, represented as thin limestone and chert breccia in central portion; best developed on southeast flank	
	White siliceous limestone	25–40
	Red and yellow shales	70–90
Total.....		130
	Unconformity	
Pennsylvanian....	Tensleep sandstone	
	Sandstone, limy, tan to gray-white in color, with small lenses and nodules of dark chert; fossiliferous	20
	Sandstone, purplish, quartzitic, steeply cross-bedded (20°–25°)	47
	Chert, and fossiliferous siliceous limestone	1
	Alternating thinly bedded and cross-bedded white sandstone, lower portion dolomitic	57
	Strongly ripple marked and cross-bedded sandstone	15
	Fine-grained sandstone often grading into pale lavender quartzite	10
Total.....		158
Pennsylvanian and Mississippian...	Amsden formation	
	Thin beds sandstone, shale, earthy limestone purplish to red and green in color	32
	Purplish shale, with interbedded earthy white to purple limestone	145
	Hard, blocky, red hematitic shale, occasional green eyes; sandstone stringers near base, one thin conglomerate layer in this material	65
	Limestone-nodular, purple pink mottled	10
Total.....		252
	Unconformity	
Mississippian.....	Madison limestone	
	Gray to blue-gray, fossiliferous, cherty limestone. Hard, brittle, thick-bedded to massive	87
	Gray, thinly bedded, softer limestone than above. Chert less abundant	138
	Breccia due collapse of post-Mississippian caves (general in area)	51
	Limestone, buff to brown, interbedded with thin gray to white limestones. Soft, granular, often ripple-marked	280
	Thinly bedded, purplish to gray limestone. Interbedded with thin layers of minutely ripple-marked, argillaceous sediment. Limestone contains large ripple-marks. Base marked by very fossiliferous, granular limestone	73
Total.....		629
	Unconformity	

TABLE 1—*Continued*

		FEET
Ordovician.....	Bighorn dolomite	
	Brown to chalky rather well-bedded dolomites, interbedded with thinly bedded gray to buff limestones, all unfossiliferous	138
	Massive, cliff forming, dull brown to chalky-buff dolomite. Poorly bedded, chert nodules in lower 75 feet, weathers with a brecciated appearance	210
	Dolomite, massive, compact, light buff to bright yellow, unfossiliferous	20
Total.....		368
	Unconformity	
Cambrian.....	Deadwood formation(?)	
	Flat-pebble limestone conglomerate, interbedded in shaly limestones, and glauconitic shales.....	253
	Glauconitic shales, with minor thinly bedded limestones, and flat-pebble conglomerate beds	500
	Basal arkosic sandstone, weathers into limonitic stained sand (Flathead quartzite)	50
Total.....		803
Pre-Cambrian....	Schists and granite	

A brief résumé of the features found in the Pryor uplift with a generalized structure-contour map and accompanying cross sections was prepared for the *Guidebook* of the Sixteenth International Geological Congress.[7] The interpretations of the structural conditions offered at that time have been modified by the writer.

STRATIGRAPHY

The stratigraphic section exposed in the Pryor Mountains represents an esentially conformable, though discontinuous, sequence from the Middle Cambrian to Upper Cretaceous—the individual formations being parts of units extensively developed in south-central Montana and north-central Wyoming. The Silurian and Devonian periods are not represented, as far as is known, in the geologic section exposed in the Pryor Mountains.

Formational names and boundaries have been adopted from the descriptions by N. H. Darton and R. D. Salisbury,[8] by Knappen

[7] W. T. Thom, Jr., N. M. Fenneman, R. T. Chamberlin, and W. H. Bucher, *Guidebook 24, 16th Internat. Geol. Cong.* (1932), pp. 47–52.

[8] *Geologic Atlas of the United States*, Bald Mountain–Dayton, Wyoming, Folio (1906).

and Moulton,[9] and by Thom, Hall, Wegemann, and Moulton[10] with little change. A later paper will deal more specifically with the stratigraphy within the area.

STRUCTURAL GEOLOGY

GENERAL

Many of the mountain uplifts and anticlines in the Central Rocky Mountains and the near-by Great Plains show certain characteristic structural forms and relationships which are also clearly and typically exhibited by the Pryor Mountain uplift. The usual surface expression of the uplifts is an asymmetric anticline, or a monocline, which traditionally has been assumed to have resulted from rotational uplift along "normal" faults. Both the steepness and the degree of asymmetry of these folds vary greatly, but the areal and structural patterns are so similar that evidently many of them may be grouped as belonging to one structural and dynamic type.

Within the Pryor Mountains it is possible to study in a compact area several of the so-called trap-door features and to investigate a sequence which includes simple asymmetric anticlines, as well as faulted ones which have been overthrust in the later stages. The sequence appears to follow a regular order of evolution in which a simple surface anticline develops into a fractured one and finally into a more complex uplift with overthrusting. Examples of asymmetric folds in various stages of folding and dissection exposed in the Bighorn Basin are the Sheep Mountain, Rattlesnake Mountain, and Grass Creek anticlines. Other examples in the general region are the Cedar Creek anticline in Montana, Lance Creek anticline in Wyoming, and the Shawmutt anticline in Montana.

Recently published tectonic maps of the Bighorn-Beartooth region[11] show graphically the distribution and predominance of the asymmetrical uplifts in this portion of the Central Rocky Mountains, of which the Pryor Mountains are a part. Bucher, in discussing this situation, has noted not only the local correspondence in

[9] *Op. cit.* [10] *Op. cit.*

[11] W. H. Bucher, W. T. Thom, Jr., and R. T. Chamberlin, "Geologic Problems of the Beartooth-Bighorn Region," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934); J. T. Fanshawe, "Structural Geology of the Wind River Canyon Area, Wyoming," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIII (1939), pp. 1442–43.

assymmetry of the major and minor features in the region but also the subdivision of the Bighorn Mountains into segments in which the directions of overturning are alternately westward and eastward. G. Duncan Johnson[12] has added detail by a study of Rattlesnake Mountain—the westward-facing, asymmetrical anticline at the south end of the Beartooth uplift. The cross section of this feature as exposed in the canyon of the Shoshone River at Shoshone Dam demonstrates clearly the relation between the surface fold and the subsurface fault.

The writer has compiled a tectonic sketch map (Fig. 1) of the Pryor Mountains and adjacent areas upon which can be seen the relation of the crustal segments discussed in this paper. It has been necessary to approximate certain contacts and positions of flexures as applied to the northern portion of the Bighorn Range as the geologic mapping in the vicinity of the Bighorn Canyon is in need of revision. The revision has been roughly sketched upon the tectonic map.

Structurally, the Pryor Mountains are divided into four major and three minor units which are similar in general structural form and vary only in detail. These units are as follows: West Pryor Mountain, Northeast Block, Shively Hill dome, Big Pryor Mountain, East Pryor Mountain, Sykes Spring area, and the secondary monocline on East Pryor Mountain. Each of the major segments has been uptilted at the northeastern corner, as the tectonic sketch map (Fig. 1) indicates and as shown in detail by the areal geologic maps (Figs. 2–6). The presence or absence of faulting, as well as the type and degree of the same, comprise the variations in structural detail.

Place names and geographic location of features are presented by means of the areal geologic maps and also upon the tectonic diagram. However, for further facility in discussion a brief description of the principal faults is presented before taking up the details of each segment.

North Pryor fault.—The fault which causes the prominent north-facing scarp and limits the uplift in that direction has been desig-

[12] "Geology of the Mountain Uplift Transected by the Shoshone Canyon, Wyoming," *Jour. Geol.*, Vol. XLII (1934), pp. 809–38.

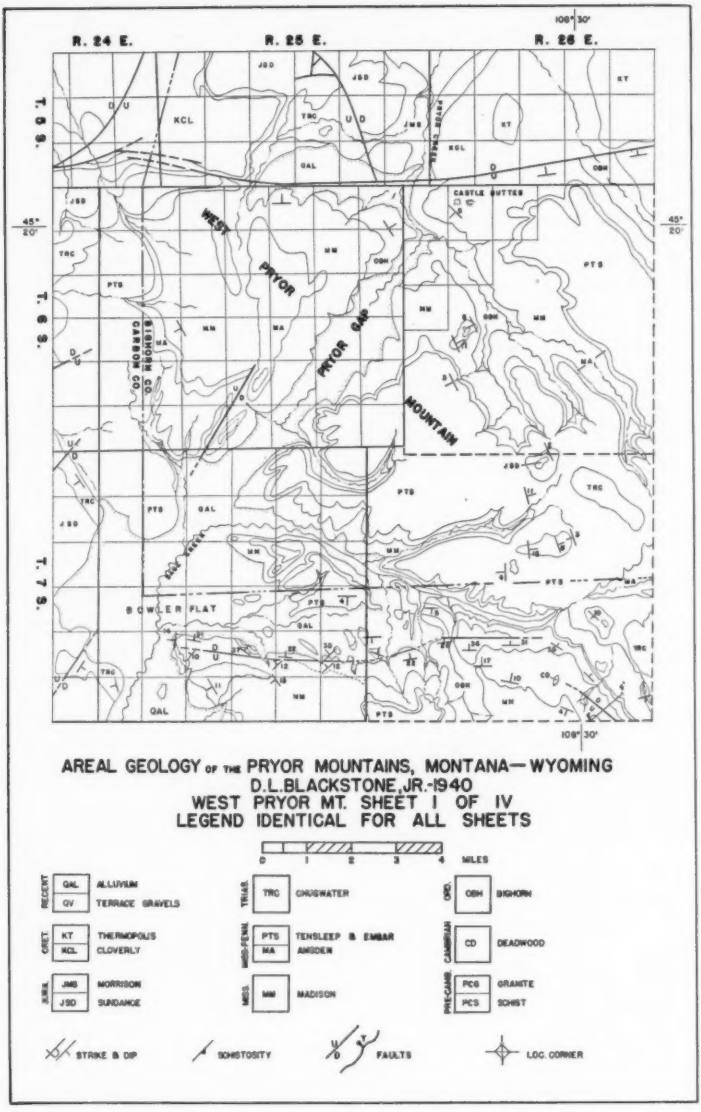

FIG. 2.—Areal geology of the Pryor Mountains (Sheet I of IV)

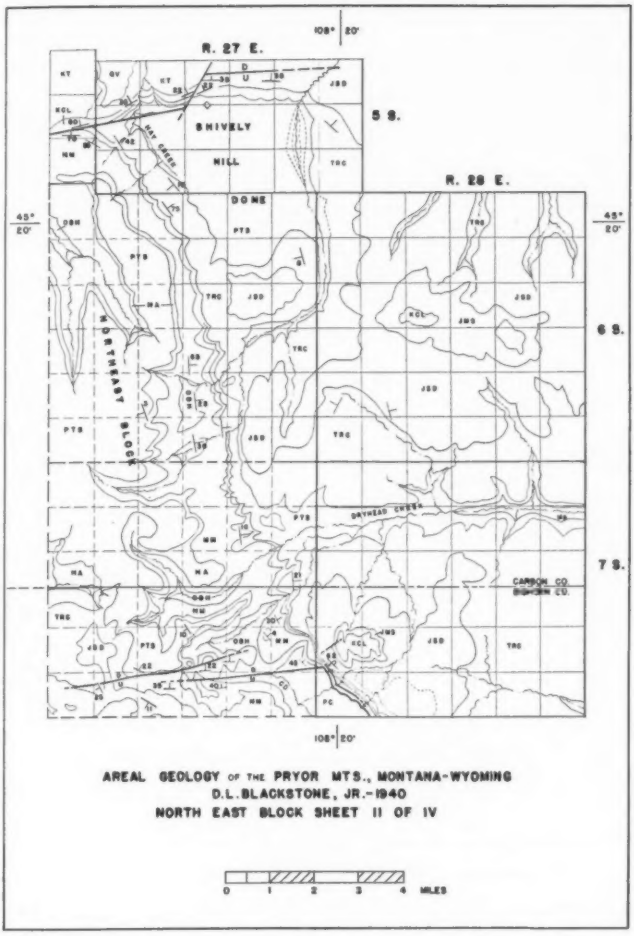

FIG. 3.—Areal geology of the Pryor Mountains (Sheet II of IV)

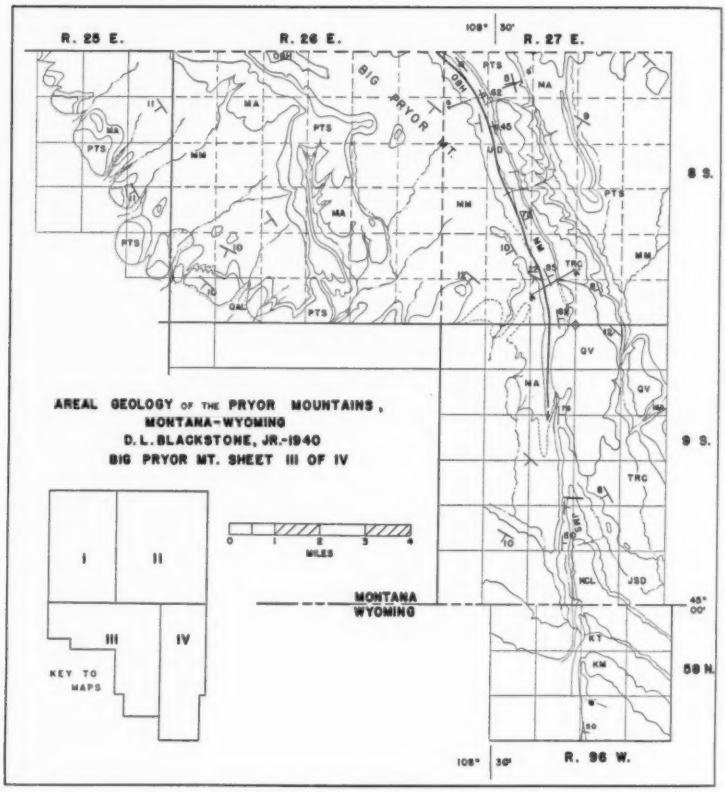

FIG. 4.—Areal geology of the Pryor Mountains (Sheet III of IV)

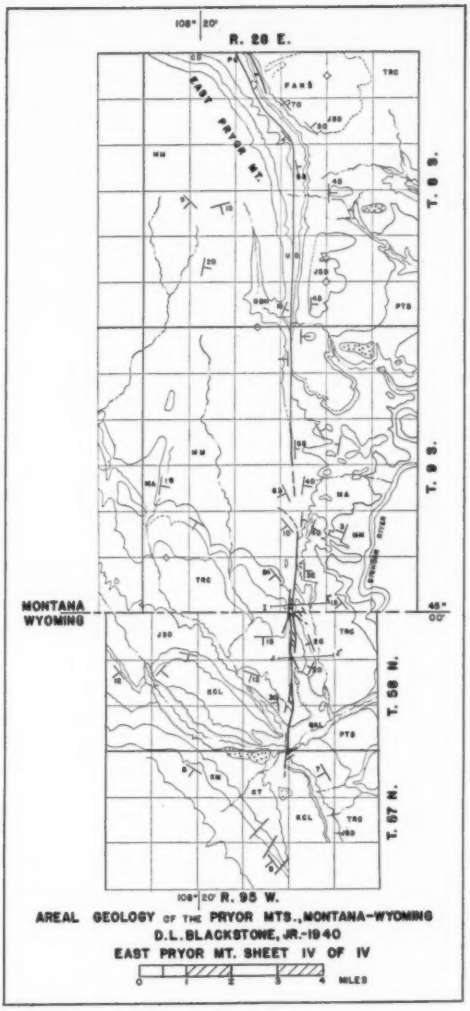

FIG. 5.—Areal geology of the Pryor Mountains (Sheet IV of IV)

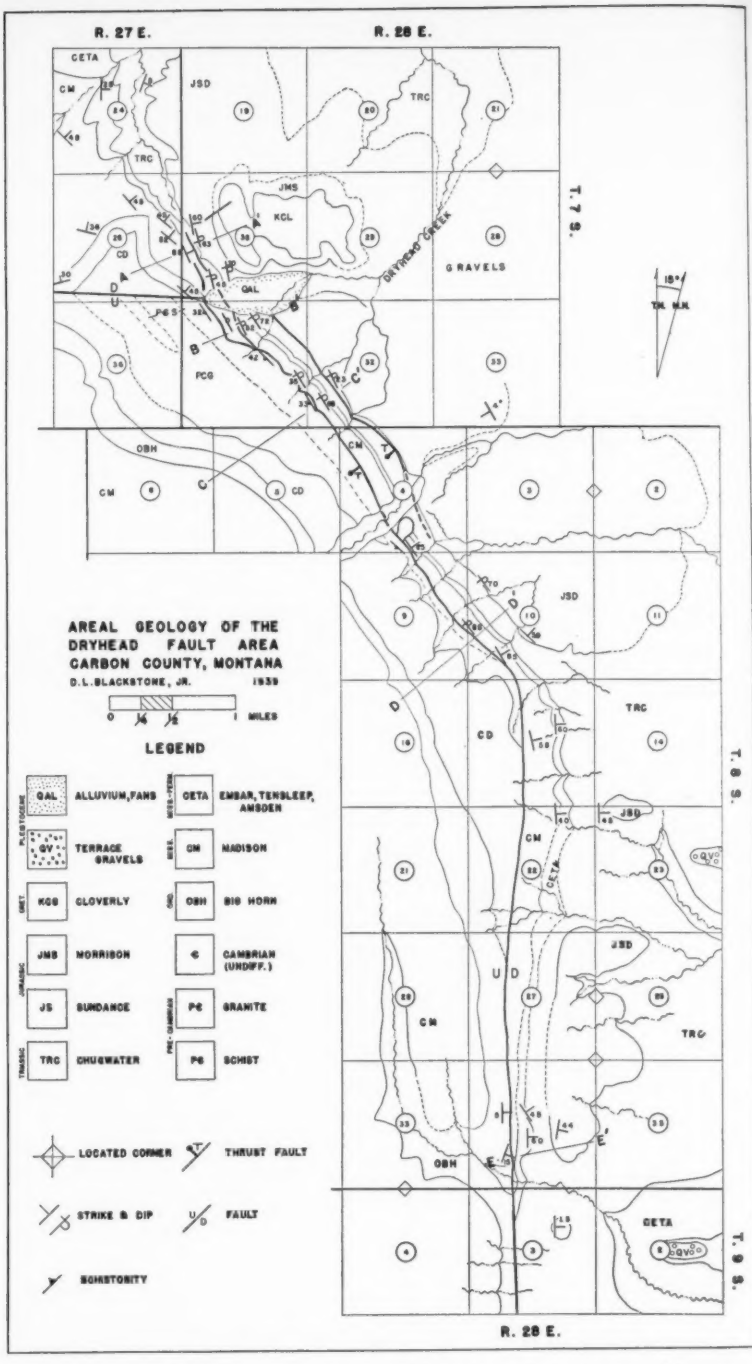

FIG. 6.—Detailed geologic map of the Dryhead fault area, Carbon County, Montana

nated the North Pryor fault by Knappen and Moulton,[13] and that designation is continued here.

Dryhead fault.—The major fault on the east side of the uplift, crossed by the headwaters of Dryhead Creek, is herein termed the Dryhead fault.

Sykes Spring fault.—A second distinct and separate surface fracture, though in the same structural line, extends southward beyond the Dryhead fault and has been designated the Sykes Spring fault because of its development in the vicinity of Sykes Spring, Wyoming.

Sage Creek fault zone.—The Sage Creek fault zone is the median east-west fault zone, which lies on the south side of Sage Creek and extends eastward from Bowler P.O. to attain maximum importance at its junction with the Dryhead fault. This zone is the eastward extension of the Nye-Bowler lineament as mapped by C. W. Wilson, Jr.[14]

Crooked Creek fault.—The fault bounding the Big Pryor Mountain block along the east side has been called the Crooked Creek fault from the stream of that name which it parallels.

A group of en echelon faults on the south flank of Sykes Mountain, which produce marked offsets in an escarpment of Mowry siliceous shale at that point, has not been named by the writer.

DETAILED STRUCTURE

The detailed mapping of the Pryor Mountains has revealed that the asymmetric features may be arranged in a sequence which demonstrates their mode of development. The stages in this sequence are as follows:

1. Simple, relatively gentle asymmetric anticlines or monoclines
2. Asymmetric anticlines with relatively high dips on the steep limb, which can be traced laterally into a ruptured fold, bounded by "normal" faults
3. Ruptured feature, bounded by a multiple fault, the lower of which is a low-angle overthrust

The specific examples which follow are believed to indicate the relation between the folding of the surface rocks and the behavior of the underlying basement rocks, the manner of fault propagation

[13] *Op. cit.*

[14] "Geology of the Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XX (1936), pp. 1168–88.

through the overlying sedimentary cover to the surface, and the nature and form of the fault surface in the upper part of the basement crystallines as reflected by the low-angle overthrusting.

SIMPLE ASYMMETRIC ANTICLINES

An excellent example of a very simple asymmetric anticline is provided by the fold bounding the West Pryor Mountain block along its eastern edge. Similar, but more strongly developed, examples are provided by the east flexure of the Northeast Block and by the secondary monocline present on the southwesterly dipping backslope of East Pryor Mountain (Figs. 1, 2, and 5).

The Tensleep sandstone which caps the West Pryor Mountain flexure rises along the flank from beneath the alluvium of Bowler Flat to the crest at dips of from 2°–4°. The sandstone descends structurally from the crest some two hundred feet at dips of 5°–12° before rising again up the dip slope of the Northeast Block at an angle which is essentially 3° SE. The Tensleep sandstone rises 1,000 ± feet to the crest of the Northeast Block before again dropping structurally in an unbroken though dissected fold to the valley of Hay Creek. The strata in this east limb strike approximately N. 30° W., and the dips range from 16° to 70° E. depending upon locality. The Tensleep along the valley of Hay Creek in the narrow syncline between the Northeast Block and Shively Hill dome is approximately two thousand feet lower in elevation than at the crest of the fold to the west (Fig. 3). The Tensleep sandstone rises again from the valley of Hay Creek to form the surface of the broad, simply asymmetrical Shively Hill dome.

The secondary monocline present upon the back slope of East Pryor Mountain trends essentially north-south, with the steeper limb facing the east. The flexure is very gentle in Sec. 18, T. 8 S., R. 28 E., but becomes more pronounced in the Crooked Creek Valley to the south. The steep limb dips approximately 20° E., the gentle west dip slope remaining that of East Pryor Mountain, namely, 5°–12° SW.

ASYMMETRIC FOLDS PASSING INTO FAULTS

Examples of the passage of an unruptured asymmetric anticline into one ruptured by what appears at the surface to be a normal fault are the Big Pryor Mountain fold, North Pryor fault zone, and

the southern portion of East Pryor Mountain. A typical case in point is provided by the Big Pryor block, which is bounded on the east by the Crooked Creek fault and on the north by the Sage Creek fault zone.

This south-plunging, asymmetric anticline as exposed along Gypsum Creek at the south end of the block merges to the north into a ruptured surficial "fold." The strata on the upthrow side of the Crooked Creek fault are essentially horizontal or dip gently to the east in the zone immediately adjacent to the fault. In contact with the fault in the relatively depressed block the strata are either steeply inclined (70°–80° E. dip) or vertical, which condition places the steeply dipping Madison limestone of the downthrow segment in fault contact with the nearly horizontal Bighorn dolomite of the upthrow side. The fault at the surface appears with one exception as a vertical normal fault, with the east segment downthrown approximately twenty-five hundred feet. The exception occurs in a limited exposure (Fig. 3), in which the fault is a high-angle reverse fault. This exposure is opposite the point of greatest elevation of the upthrown mountain mass.

For a distance of approximately a mile in Sec. 6, T. 8 S., R. 27 E., the Madison limestone in the depressed segment has been overturned as much as 45° so that the strike now is N. 60° W., and the dip is 45°–62° W. Within this area the dip of the fault plane, though poorly exposed, must be at least as low as the angle indicated by the dip of the overturned strata. This overturning is symptomatic of the next stage in the development of these anticlines.

The reader is here referred to an excellent photograph which illustrates a similar fold in a more advanced stage of dissection but slightly less well developed structurally. The Dry Fork anticline of the northern Bighorn Mountains as figured by Darton[15] and Thom,[16] is an asymmetric fold in which the controlling fault in the basement is exposed, yet it has not been propagated upward through the entire overlying section but has fractured only the Cambrian (Deadwood?) strata.

The uptilted segment of the Big Pryor Mountain does not directly

[15] *Op. cit.*, Pl. 41 *B*, p. 98.

[16] W. T. Thom, Jr., *et al.*, "Geology of Big Horn County and the Crow Indian Reservation, Montana," *U.S. Geol. Surv. Bull. 856* (1935), Pl. 7.

reveal the character of the movement within the underlying crystallines but nonetheless fits the hypothesis herein to be presented. The fault, which is postulated beneath both the West Pryor and the Northeast blocks, has here ruptured the sediments of the surface fold, where a differential of 2,500–3,000 feet of elevation has taken place and in so doing has given rise to a suggestion of incipient thrusting as a forerunner of events to come.

RUPTURED ASYMMETRIC ANTICLINE WITH SUBSIDIARY THRUSTING

The uplifted segment, East Pryor Mountain, has been elevated and rotated more than any other segment within the mountain mass and has therefore been more fractured along the margin. The fault movement has taken place on the Dryhead fault and on the eastern portion of the Sage Creek fault zone, thus relieving stress in two directions. This large segment is fundamentally similar to Big Pryor Mountain but has passed beyond the stage of a simple "normal" fault to a condition wherein the downthrow block has been modified by low-angle thrust faults.

The details of this faulting are presented by a map, cross sections (Figs. 6 and 7), and photographs (Figs. 11–15). The Dryhead fault plane is moderately well exposed and can be readily followed for a part of its length as the contact between the Paleozoic sediments and the basement rocks. The location of this fracture has been influenced by an older line of weakness, as is shown in the close agreement between the attitude of the schistosity and the dip of the fault plane. The schist (hornblende-mica) as exposed is actually only a narrow zone bounded on one side by the Dryhead fault and on the other by the relatively sharp contact between a gray and a pink granite, intrusive into the schist. This schist-granite contact has undoubtedly been a controlling factor in localizing the crustal failure. Incipient thrusting has taken place on the primary Dryhead fault so that locally the pre-Cambrian rests on the Mississippian Madison limestone, but the dominant lateral movement has taken place because of the development of the low-angle shear planes.

At the southern exposures of the primary fracture the surface displacement is relatively small, and the fault may be classified as a normal fault which is approximately vertical. As the observer proceeds north along the fault, the dip of the fault plane changes from

vertical to a westward dip and becomes less steep, so that the fault that is first observed as a normal fault grades into a high-angle reverse and later to a low-angle reverse fault, if 45° be used as the limiting case between the two. In addition to this primary fracture

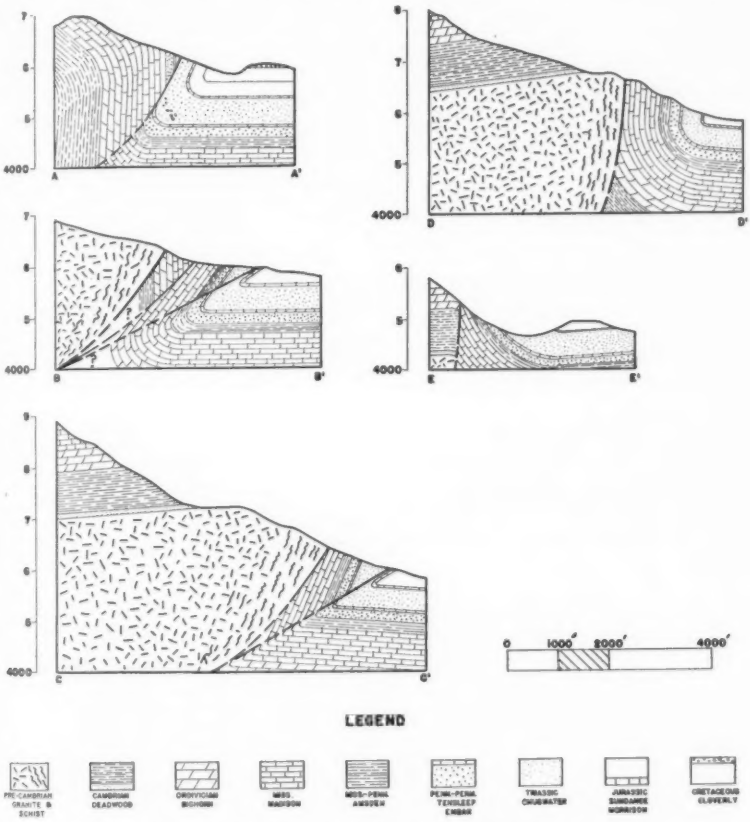

FIG. 7.—Geologic cross sections to accompany Fig. 6

there are secondary subsidiary thrust faults which lie beneath the Dryhead fault, outcrop to the east of it, and have a much lower angle of dip as is shown both by the exposure of the fault plane and by the reversed dips in the strata beneath. It is this subidiary shear

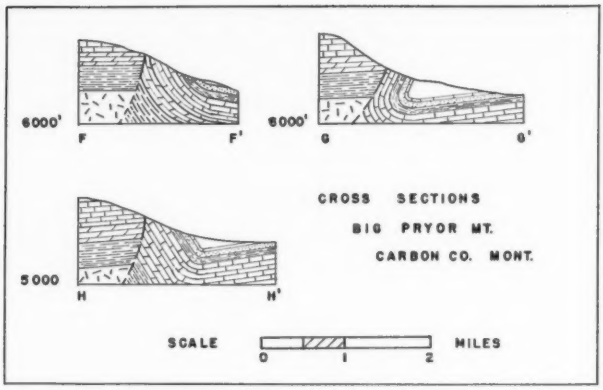

FIG. 8.—Geologic cross sections of Big Pryor Mountain

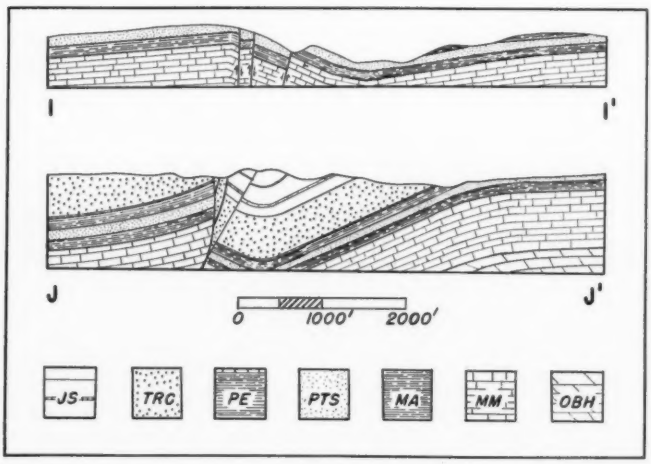

FIG. 9.—Geologic cross section of the Sykes Spring area, Wyoming

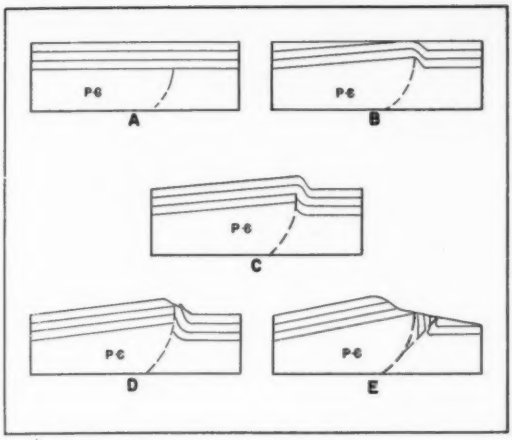

FIG. 10.—Cross sections demonstrating relation of subsurface fracturing to surface folding.

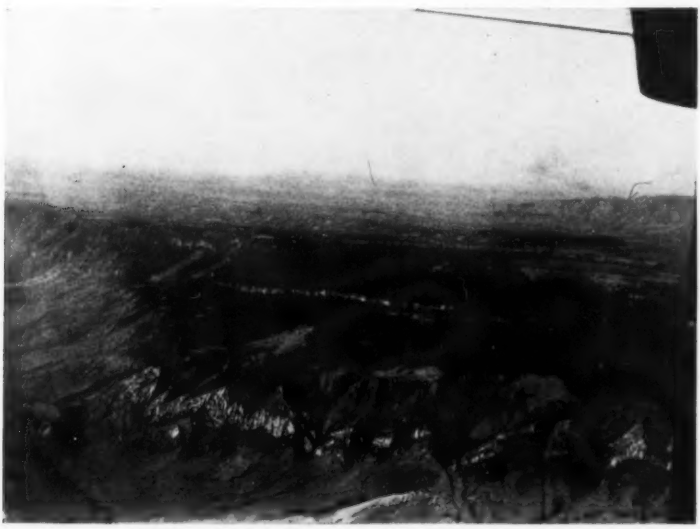

FIG. 11.—Aerial view southwest across the scarp of East Pryor Mountain from an elevation of 12,000 feet.

FIG. 12.—View across Layout Creek Canyon. Cambrian strata in fault contact with Madison limestone.

FIG. 13.—View north from a point in Sec. 9, T. 8 S., R. 28 E., Montana. Pre-Cambrian in fault contact with Madison limestone.

FIG. 14.—View south across the Sykes Spring fault zone

FIG. 15.—View north showing the Dryhead fault and subsidiary thrust beneath it as exposed on the headwaters of Dryhead Creek.

that makes possible the low-angle thrusting of the previously elevated segment of the granitic basement.

Another example of a ruptured anticline incipiently passing to overthrusting is present in the Sykes Spring area in the extreme southeast portion of the map area (Fig. 9). The surficial fracture is separate from the Dryhead fault, yet lies in a direct south projection of that fracture and may well be considered as controlled by a common zone of weakness in the basement rocks. Within this area the potentially east-moving segment of the Pryors impinges upon an opposite tendency (west-facing asymmetry) as expressed by the Porcupine Creek anticline that extends northwestward from the main range of the Bighorns. The dominant movement has been from west to east, with the resulting faulting and overriding of the intervening syncline in that direction by the Pryor Mountain segment.

STRUCTURAL SIGNIFICANCE OF THE PRYOR RELATIONSHIPS

The evidence as presented by the areal geology and the geologic cross sections of the faults bounding the Pryor Mountain segments leads the writer to the conclusion that the theory of normal-fault origin for the asymmetric anticlines as proposed by Thom and by Thom *et al.*, and by Johnson[17] requires modification. In view of the overthrusting developed within the area mapped by the writer and because of the similarity of these features to those examined on the west flank of the Bighorn Mountains in the vicinity of Cookstove Basin and because of conditions shown in Cross Section 1, accompanying Johnson's map of Rattlesnake Mountain, the hypothesis of control by normal faults is here replaced by a dynamically different interpretation. The interpretation placed by the writer on the phenomena observed is as follows. The Pryor Mountain area in response to the influence of regional tangential pressure from the southwest during the Laramide orogeny has behaved somewhat as shown diagrammatically in Figure 10. Pre-existing zones of weakness which have been substantiated for many parts of the region[18] appear to

[17] Thom, *op. cit.;* W. T. Thom, Jr., *et al.*, "Yellowstone-Beartooth-Bighorn Region," *Guidebook 24, 16th Internat. Geol. Cong.* (1933); Johnson, *op. cit.*, pp. 809–37.

[18] Ernst and Hans Cloos, "Pre-Cambrian Structures in the Beartooth, Bighorn, and Black Hills Uplifts, and Its Coincidence with Tertiary Uplifting," *Proc. Geol. Soc. Amer.*

have localized failure of the crust, with differential elevation resulting along the fractures. For reasons to follow, the surfaces upon which the differential elevation took place are believed to have a westward dip and to be curved surfaces. The result of this initial movement (*B* and *C*) may or may not have resulted in a surface flexure, as the total thickness of the section is a variable to be considered. A relatively thick section might not be affected at the surface. Renewed movement on the same plane will definitely result in a surface flexure (*C*), asymmetric in character, in the present case with a steep east limb.

Further response to the tangential pressure results in the projection of the basement fracture upward through the sedimentary cover to the surface (*D*). The surface feature is now an asymmetric anticline fractured by what may be described as a "normal" fault nearly perpendicular in attitude.

Carried on further this rotation and elevation of the segment no longer responds to lateral forces by rotation and essentially vertical release from stress. A relative disengagement of masses of basement rocks upon the opposite sides of the fracture is achieved in this stage, following which a new failure occurs. This secondary failure takes the form of a low-angle shear plane (*E*), the purpose of which appears to be a lowering of the attitude of the initial fracture surface in order that lateral movement may follow with less expenditure of force in further elevation. The low-angle shear breaks through the corner of the depressed block at an angle observed in the field to be as low as 40° and inferred from the attitude of overturned strata to be as low as 23°.

The above explanation has been adopted to account for the relations which can be seen at East Pryor Mountain and elsewhere. The indropped wedge and wide belt of overturned strata depicted by Johnson[19] do not seem explicable by the normal gravity faults which he uses in his cross section. The overturned beds might better

(1934), p. 56; C. W. Wilson, Jr., "A Study of Jointing in the Five Springs Area, East of Kane, Wyoming," *Jour. Geol.*, Vol. XLII (1934), pp. 498–522; E. C. H. Lammers, Beartooth Mountains, personal communications.

[19] *Op. cit.*, pp. 810–11, Structure section No. 2.

be explained as the result of a low-angle thrust in approximately the position of the second or outer normal fault.

The writer has in reconnaissance examined localities on the west flank of the Bighorn Mountains in the vicinity of Cookstove Basin, located on the Dayton–Bald Mountain Folio.[20] In these localities the writer observed a high-angle, early fracture upon which vertical movement had occurred. At a lower elevation there are low-angle shear planes not shown on the Folio map, upon which incipient thrusting has taken place. The observed conditions can be explained by the above hypothesis but do not fit the normal gravity-fault explanation.

Two recent wells drilled at Lance Creek field in Wyoming may be interpreted along similar lines. The Argo Oil Corporation Ong No. 1, SE. SW. Sec. 31, T. 36 N., R. 65 W. (*Oil Weekly*, January 22, 1940, p. 54), passed through Triassic red beds and into black Cretaceous shales above the Dakota sandstone at 5,132 feet. The test drilled black shales to 5,623 feet, at which depth the strata were essentially vertical. A second well in the same field passed from Pennsylvanian strata back into Cretaceous shales, probably in the Thermopolis formation. These wells are on the steep side of an asymmetric surface anticline, and the formational relations can hardly be interpreted as other than the result of a reverse fault which has controlled the development of the surface anticline.

A. A. Baker[21] has discussed the geology of several large monoclines in southeastern Utah. He similarly suggests that they may be controlled by curved fractures in the basement rocks which have responded to regional compressional stress. These monoclines of southeastern Utah are larger features than the ones described in the Pryor Mountains but appear to be explicable by the same mode of origin.

The presence of overturned strata adjacent to the fault in the downthrow block in many of the asymmetric anticlines is the feature most difficult to explain by normal faulting. It seems improbable that such overturning can result unless later movement has violently disturbed the original fracture, causing a reversal in the direction of

[20] Darton and Salisbury, *op. cit.*

[21] "Geologic Structure of Southeastern Utah," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIX (1935), pp. 1472–1507.

movement. If, however, the initial fracture curves beneath the upthrown segment, no added condition such as reversal of movement is necessary.

RELATION OF VERTICAL UPLIFT TO OVERTHRUSTING

The asymmetric anticlines within the Pryor Mountains are more sharply folded (or have the greatest fault displacement) opposite the points of greatest elevation than elsewhere. Following the postulate of origin by tangential pressure acting on possibly curved fault surfaces dipping beneath the upthrown segments, there appears to be a definite relationship between the amount of vertical uplift and the amount of overthrusting.

The Northeast Block is elevated approximately 2,000 feet with no faulting, whereas the Big Pryor segment has been elevated approximately 2,500 feet and is fractured by an incipient thrust fault. East Pryor Mountain has been elevated a minimum of 2,600 feet, and the thrusting is well developed. In the Five Springs Canyon area in the Bighorn Mountains C. W. Wilson[22] indicates a vertical uplift of approximately one mile, but owing to the relief of stress upon fractures and joints in the granite the thrusting is less than might be expected, only 600 feet of horizontal movement being recorded.

West of Buffalo, Wyoming, as mapped upon the Cloud Peak–Fort McKinney Folio, overthrusting of the pre-Cambrian core of the range upon the Cretaceous sediments is of the order of magnitude of $2\frac{1}{2}$ miles. The thrusting is due east of the maximum elevation of the range, 25,000 feet above the adjacent basin areas, as given by Darton.[23]

The Beartooth Mountains present another example of thrusting of great magnitude correlated with maximum vertical elevation. The thrusting eastward in the Beartooth Mountains is greater than ten miles, with a relative vertical elevation of approximately 35,000 feet.[24] The maximum thrusting is closely related in space to the maximum vertical uplift.

[22] "A Study of the Jointing in the Five Springs Creek Area, East of Kane, Wyoming," *Jour. Geol.*, Vol. XLII (1934), p. 501.

[23] *Op. cit.*, p. 91.

[24] W. T. Thom, Jr., O. T. Jones, and R. T. Chamberlin, "Morphology of the Beartooth Uplift, Montana and Wyoming," (abst.), *Proc. Geol. Soc. Amer.* (1935), p. 117.

SIMILARITY OF PRYOR STRUCTURE TO RAMP STRUCTURE

The structural trough lying between the northern portion of the Bighorn Mountains and the east flank of the Pryor Mountains is known as the Dryhead-Garvan Basin (Fig. 1). The vertical-walled, deeply entrenched, meandering valley of the Bighorn River in its exit from the Bighorn Basin divides the single structural trough into two geographic units of which the Garvan Basin lies southeast of the river. One boundary of the elongate trough is the west-facing asymmetric faulted fold of the northern Bighorn segment. The crest of the range is broad and flat in the northern portion, hence the steeply dipping beds of the flank are very prominent. The western margin of the trough is the Pryor Mountain uplift that has been described in this paper. The Pryor trend is potentially an eastward-thrusting mass, while that portion of the Bighorn Range north of the Black Gulch fault tends to move in the opposite direction—to the west. The trough between uplifted segments is similar in character to the ramp structures described by Bailey Willis[25] in the Dead Sea area. The faults postulated for the Pryor uplift are like those suggested by Willis, i.e., reverse faults with curved surfaces, activated by tangential pressure, and emerging at the surface essentially as high-angle normal faults. The upper surface of the uplifted segments is in both cases tilted away from the fault.

C. W. Washburne[26] has noted the relation of surficial folds to underlying faults and has suggested that the surface of a fault is concave toward the more active side; i.e., concave toward the downthrown side in normal faults and toward the upthrown side in reverse faults. Washburne believes that most normal faults are due to compression, with the footwall the more active segment. The writer follows Washburne in the nature of curvature of the fault plane, in this case concave toward the upthrown, active side.

ANALYSIS OF THE REGIONAL FAULT PATTERN IN THE PRYOR UPLIFT

Three principal trends control the form of the Pryor Mountains. The east-west trend of the North Pryor fault and the Sage Creek

[25] "Dead Sea Problem: Rift Valley or Ramp Valley?" *Bull. Geol. Soc. Amer.*, Vol. XXXIX (1928), p. 501.

[26] "The Origin of the Normal Faults," *Jour. Wash. Acad. Sci.*, Vol. XVIII (1928), p. 346.

fault zone are transverse to the long axis of the uplift as well as to the trend of the marginal features in the northeast portion of the Bighorn Basin. The Sage Creek fault zone as previously mentioned is an en echelon continuation of the Nye-Bowler lineament known to be a regional feature of first-order magnitude. The North Pryor fault extends 18 miles along the north flank of the uplift and is possibly related in origin to the Nye-Bowler trend.

The trend of the long axis of the uplift is composite, however, and consists of two separate directions of trend. In the southern half of the uplift the folds and faults trend north-south. Near the south edge of T. 7 S., Montana, the trends change direction to N. 30° W., and so continue to the north edge of the mountains.

The idea has been advanced that the present form of some of the features in the Yellowstone-Beartooth-Bighorn region is controlled by zones of weakness in the pre-Cambrian basement which predate the Laramide orogeny. The evidence accumulated in such cases as the Nye-Bowler lineament, the Tensleep fault, and the Lake Basin fault zone substantiates the idea. In the case of the Pryor Mountains the writer believes the same condition exists. One reason for this conclusion is the fact that the Sage Creek fault zone is a part of the larger Nye-Bowler feature. Another reason is the close correlation between the attitude of the schistosity in the zone of pre-Cambrian schist exposed on East Pryor Mountain and the attitude and position of the plane of the Dryhead fault.

In 1933 Drs. Ernst and Hans Cloos[27] concluded after a study of the pre-Cambrian structure of the Beartooth, Bighorn, and Black Hills uplifts that "though differing individually in every detail, the three structural units seem to indicate coincidence between pre-Cambrian doming and their present boundaries." If such coincidence exists elsewhere in the region it is to be expected in the case of the Pryor uplift.

R. T. Chamberlin[28] has recently pointed out the influence of the pre-Cambrian rocks and structures upon the deformation around the Bighorn Basin.

[27] *Op. cit.*

[28] "Diastrophic Behaviour around the Bighorn Basin," *Bull. Geol. Soc. Amer.*, Vol. L (1939), p. 1903.

The deformation in the Pryor Mountain uplift has resulted from the Laramide orogeny but cannot be defined more closely because of the lack of late Cretaceous or post-Cretaceous strata that might be used for dating purposes.

SUMMARY AND CONCLUSIONS

The writer believes that the studies of the Pryor uplift lead to the following conclusions.

1. The Pryor blocks are featured by median, as well as by marginal, depressions and cannot be assumed to have resulted from epanticlinal faults resulting from the upbulging of material at a center in consequence of compression.

2. The initial asymmetric anticlines of the uplift grade laterally into visible "normal" faults along the steep edges of the uptilted blocks, and these in turn grade (at least in three cases) into double faults with an "indropped" wedge between, which is really a fault slice bounded by low-angle thrusts. The outermost of these faults becomes lower and lower in angle as the height of the uplifted block increases, until the overriding mass is thrust outward on a plane which dips to the west at an angle of 23° or less as shown by the drag in the beds beneath the thrust fault.

3. In the individual segments the initial release from tangential stress has been by rotation and vertical uplift, with the later relief taking place on the low-angle shear planes which develop as a subsidiary feature. The rotational and eastward movements are believed to have taken place upon a fault plane that curves downward beneath the upraised segment.

ACKNOWLEDGMENTS.—The writer wishes to express his thanks to Drs. Rollin T. Chamberlin, Walter H. Bucher, and W. T. Thom, Jr., for their valuable assistance and suggestions during field work. The writer is especially indebted to Dr. W. T. Thom, Jr., for suggesting the area for investigation and for his continued aid during the progress of the work. He is also indebted to the following men who have aided him in various capacities in the field: Freeman Foote, Dr. John Lucke, Wallace McIntosh, Eric Bucher, and William T. Pecora, II.

The writer also wishes to acknowledge the financial support of Princeton University and the Yellowstone-Bighorn-Beartooth Research Project.

STRATIGRAPHY AND CORRELATIONS IN THE BIRCH–SLATE LAKES DISTRICT, NORTHWESTERN ONTARIO

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ABSTRACT

The stratigraphic succession of part of the Archean volcanic and sedimentary assemblage in the Birch–Slate Lakes district does not conform to the simple Keewatin-Timiskaming concept of an earlier dominant volcanic period followed by a period of sedimentation. Sedimentary formations of the Slate Lake series, formerly considered of Timiskaming age, occur with structural unconformity stratigraphically beneath the so-called Keewatin volcanics which are, in turn, overlain by younger infolded sediments. A knowledge of the stratigraphic succession of the rock units in this district permits correlations of the different series with those in adjoining areas.

INTRODUCTION

The Birch–Slate Lakes district is situated in the Patricia portion of northwestern Ontario approximately 65–100 miles north-northwest of the town of Sioux Lookout. The region is underlain by an assemblage of early pre-Cambrian volcanic and sedimentary rocks that has been traced intermittently eastward from Lake Winnipeg to where the pre-Cambrian formations are overlain by flat-lying Paleozoic sediments south of Hudson Bay. This schist belt is bounded on the north and south by younger intrusive granites and is effectively separated from other east-west trending belts of volcanic and sedimentary schists by broad areas of granite and gneissoid granites. Most of the geological studies along the belt up to the present time have been concentrated in the 250-mile-long section that extends eastward from the Rice Lake and Oiseau River districts of Manitoba, into Ontario through the valley of the English River and to the north through the basin of Red Lake, through the Birch–Slate Lakes district and eastward to Lake St. Joseph. The rocks of the southern part of the Rice Lake–Lake St. Joseph belt have been referred by Pettijohn[1] to the Rice Lake–Miniss series.

In the earlier reports of the Geological Survey of Canada the pre-

[1] F. J. Pettijohn, "Early Pre-Cambrian Geology and Correlational Problems of the Northern Subprovince of the Lake Superior Region," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 153–202.

granite rock formations of this part of Ontario were collectively considered Huronian (and later Keewatin) in age. As geological work proceeded in the region, there was a growing recognition that a considerable part of the terrain is underlain by rocks of sedimentary origin, and a place was made for these rocks in the geologic time scale above or below the volcanic formations, depending on their stratigraphic relation to them, if established for a particular area. The fact that such a relationship has in so few instances been established, while terms with time significance have been applied indiscriminately to different volcanic and sedimentary series on the basis of assumed relationships, has resulted in a confused nomenclature, and the true stratigraphic succession of some districts is only now becoming apparent.

The Rice Lake–Lake St. Joseph schist belt is bounded on the north by granite and granite-gneiss, most of which represents batholithic areas. To the south of the belt there is an extent of 20–50 miles of granite, south of which lies a number of folded belts of sediments and volcanics situated between the northern line of the Canadian National Railway and the international boundary, and including the Lake of the Woods and Rainy Lake districts. Within this belt, as in the case of others, the fold axes are generally parallel to the trend of the belt; and, as a result of this feature, correlations, to a limited extent, are possible within the belt. But, owing to the intervening areas of plutonic rocks between the belts, satisfactory correlations cannot be made from one belt to another in the light of our present knowledge. The southern part of the Rice Lake–Lake St. Joseph belt is underlain by dominantly sedimentary rock types, and the northern part by predominantly volcanic rocks that contain thick, limited lenses of younger infolded sediments.

The present paper is based largely on a detailed study of the Uchi–Slate Lakes map area embracing approximately 170 square miles and situated roughly midway between Lake St. Joseph and the Ontario-Manitoba boundary (Fig. 1). The boundary between the principal volcanic and sedimentary series crosses the map area (Fig. 2). The work was done in connection with a field study for the Ontario Department of Mines, and the writer is indebted to Dr. M. E. Hurst, provincial geologist, for permission to publish the results. The attempt to carry correlations into adjoining areas would

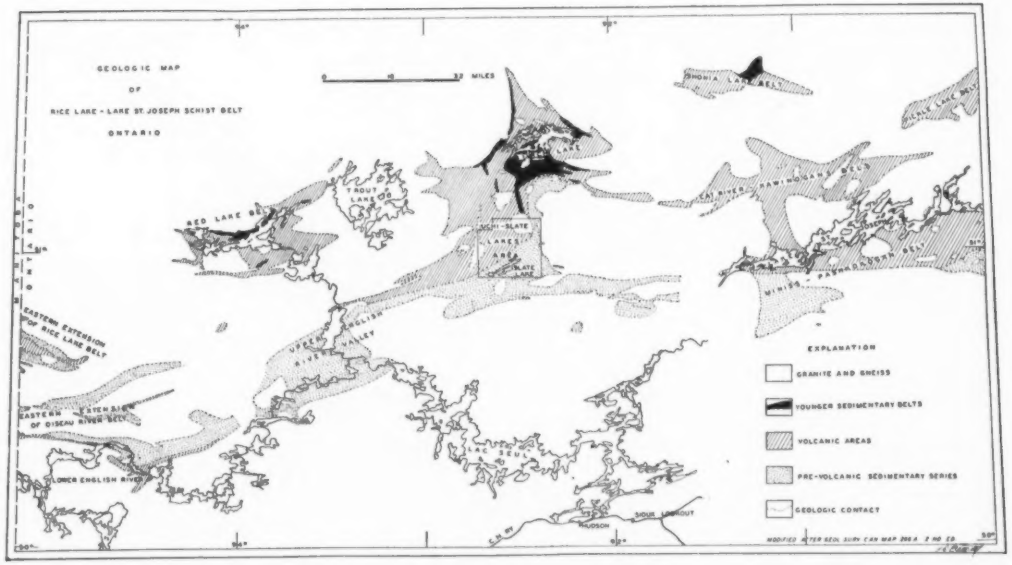
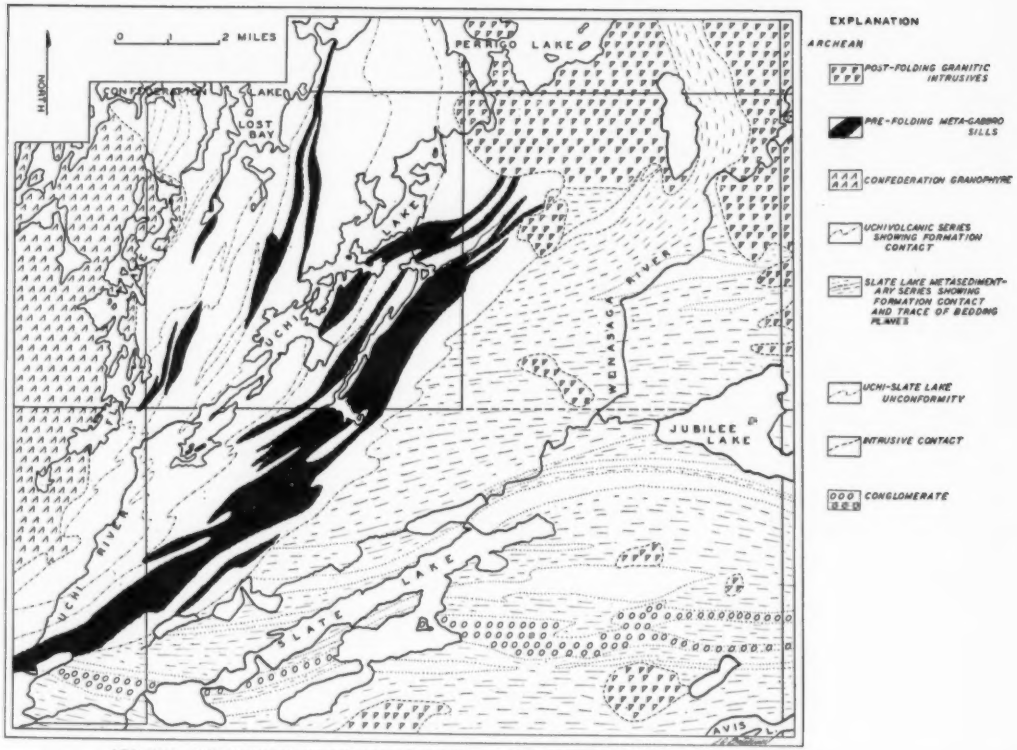

GEOLOGIC MAP
OF
RICE LAKE - LAKE ST. JOSEPH SCHIST BELT
ONTARIO
0
10
20 MILES
INDIA LAKE BELT
TROUT LAKE
RED LAKE BELT
SUSH-SLATE LAKES AREA
SLATE LAKE
CAT RIVER
WABIGOON BELT
MINAKI
EASTERN EXTENSION OF RICE LAKE BELT
EASTERN REMNANT OF OISEAU RIVER BELT
ENGLISH RIVER
LAKE SEUL
LOWER ENGLISH RIVER
SOUR LONGBOUT
RAWSON
EXPLANATION
GRANITE AND GNEISS
YOUNGER SEDIMENTARY BELTS
VOLCANIC AREAS
PRE-VOLCANIC SEDIMENTARY SERIES
GEOLOGIC CONTACT
MODIFIED AFTER GEOL SURV CAN MAP 306 A 2 ND ED

FIG. 1



GEOLOGIC SKETCH MAP OF UCHI-SLATE LAKES AREA NORTHWESTERN ONTARIO

FIG. 2

not have been possible without the extensive studies made in the region during the last few years by geologists of the Ontario Department of Mines.

PROBLEMS OF PRE-CAMBRIAN CORRELATIONS

BASIS OF CORRELATION

The early pre-Cambrian terrain abounds with structural complexities that only the most detailed studies can fathom, and, as a consequence, the proper interpretation of Archean stratigraphy is dependent on the solution of structural problems. The most practicable method of broad-scale correlation is based, first, on the determination of the sequence of the predominantly sedimentary and predominantly volcanic groups and, second, on the tracing of these groups into adjacent areas along the strike of the formations. In this way correlations of a tentative nature may be made as far as the stratigraphic sequence of the different groups remains unchanged. The greatest limitation to the application of this method is that, owing to the continental origin of much of the Archean, the rock units have a limited lateral extent, and individual formations cannot ordinarily be traced for any great distance. As a result of the alternation of volcanism with sedimentation, together with simultaneous volcanic and sedimentary deposition in different districts, a sedimentary group may pass laterally into a volcanic group of the same age. As this is one of the greatest stratigraphic hurdles at the present time, correlations are only reliable in so far as a similar lithologic sequence may be traced.

In Archean rocks the use of comparative criteria for the determinations of age relationships, such as the relative amount of deformation and the degree of metamorphism, has no widespread application, except in so far as these features may be related to specific orogenies. Unconformities also are commonly of only local value in making correlations, and many ancient pre-Cambrian conglomerates, instead of representing a widespread unconformity, are merely accumulations of coarse detritus (often in the form of long lenses) in response to local uplift, or deposited along a downfaulted block. The relation of intrusives to stratified rocks is of little help as the intrusives themselves cannot be correlated, except within limited areas.

GEOLOGIC SKETCH MAP OF UCHI-SLATE LAKES AREA NORTHWESTERN ONTARIO

FIG. 2

Periods of granitic intrusion concomitant with orogenesis form an ideal basis for subdivision of the pre-Cambrian but are hardly practicable in the subdivision of the Archean. In the region northwest of Lake Superior, as in many other parts of the Canadian Shield, the principal period of granitic intrusion followed the folding of the volcanic and sedimentary formations. Except in a few isolated districts, more than one age of granite cannot be proven in place, although an earlier granite is often assumed on the basis of granitic pebbles in conglomerates. Earlier deformations are, in most instances, effectively masked by later ones. Studies in any district show that the granites, with few exceptions, are intrusive into the folded schists. It is for this reason that age determinations based on studies of radioactive minerals are of little value for making comparative correlations of the older rocks, except that they indicate the order of magnitude. Radioactive minerals are for the most part found associated with pegmatites spatially (if not genetically) related to near-by granites, and, although their study may yield the approximate age of intrusion, the results cannot ordinarily be applied to compare the ages of the invaded schists and gneisses. It is possible that in some of the schist belts folding and granitic intrusion have been slowly progressive from one end to the other, possibly over extensive geologic periods. If this were so, the sediments and volcanics depositing at one end of a belt would overlap in time the intrusion of granites near the other end. Correlations from one schist belt to another cannot be made with any security, as the intervening granitic rocks cannot be bridged except in the realm of speculation.

PRE-CAMBRIAN NOMENCLATURE

Not only does the basis of correlation in pre-Cambrian rocks present a problem but the nomenclature used is subject to so many interpretations that unfortunate misunderstandings have often arisen. This chaotic condition is principally the result of the indiscriminate application of series and time terms based on the fragile criterion of lithologic comparison. Thus the tendency to apply the term "Keewatin" to most early pre-Cambrian volcanic rocks presupposes a time correlation that, in most cases, cannot be proven. Similarly there has been a practice of giving the name "Timiska-

ming" to many early pre-Cambrian sedimentary groups in cases where the age of the sediments relative to the so-called Keewatin volcanics was not established. Even the disputed "Coutchiching" has reappeared as a series term in areas of mica schists that may be the result of igneous metamorphism of impure quartzite or slate, no matter what their age. Thus by usage, the classifications formerly constructed to conform as far as possible to a time scale have degenerated to a condition of lithologic application. As a result, the pre-Cambrian stratigraphy of many regions is meaningless.

The history and the development of pre-Cambrian terms are not within the scope of this paper, and most of the terms are fully discussed in two carefully documented papers presented before the Royal Society of Canada by G. A. Young,[2] chief geologist of the Geological Survey of Canada. In addition, the late R. W. Brock[3] has given a valuable interpretation of some of the earlier pre-Cambrian classifications.

In northwestern Ontario we find that the volcanic and sedimentary rocks comprising the different schist areas have all been involved in complex folding, which in any district has been followed by granitic intrusion. Is this the close of Archean igneous activity? Huronian and Keweenawan formations in Ontario and other parts of the Canadian Shield are relatively flat-lying and little disturbed, except where they have been involved in mountain-building movements, such as the folded belt along the north shore of Lake Huron. No late pre-Cambrian orogenic belts have been recognized in Patricia, and in the entire region northwest of Lake Superior the widespread younger granites and older formations they intrude show little evidence of having been disturbed by postgranite mountain-building movements. If these granites have marked the close of the Archean, the widespread sedimentary strata of northwestern Ontario are of pre-Huronian age.

An interesting light on the relative order of magnitude of the age

[2] "The Geological Investigation of the Canadian Shield (Canadian Portion), Part I: 1882–1897," *Trans. Roy. Soc. Canada*, Vol. XXVI, Sec. 4 (1932), pp. 341–71; "The Geological Investigation of the Canadian Shield (Canadian Portion), Part II: 1897–1912," *ibid.*, Vol. XXVII, Sec. 4 (1933), pp. 67–108.

[3] "Notes on the Pre-Cambrian of the Canadian Shield with Reference to Pre-Cambrian Nomenclature," *Geol. Mag.*, Vol. LXXIII (1936), pp. 119–41.

of the granitic intrusions is given by the lead-method determinations from radioactive minerals associated with granites occurring near Sudbury in the Huronian area and south of the Rice Lake district of Manitoba. The ratio Pb/U + 0.36 Th from the Huronian area granite[4] indicates an age between 740 and 840 million years. On the other hand, De Lury and Ellsworth have determined the lead ratio from uraninite[5] occurring in the Oiseau River district north of the Winnipeg River near the western end of the Rice Lake–Lake St. Joseph Belt, the result indicating a vastly greater age than that determined for the age of a granite in the Huronian area. The ratio Pb/U + 0.38 Th = 0.260, according to Adolph Knopf,[6] represents an antiquity of 1,620 million years. The uraninite occurs in pegmatites said to be related to a granite that intrudes the Rice Lake series. As this is one of the greatest ages reported from lead determinations in North America, the Rice Lake series is thus placed among the most ancient pre-Cambrian rocks so far recognized not only in the Canadian Shield but throughout the world. A similar place in the time scale is indicated as far as correlations can be carried eastward along the Rice Lake–Lake St. Joseph schist belt.

Table 1 shows the pre-Cambrian classifications of Logan, Miller, Leith, Lawson, and Collins, together with one applying to northwestern Ontario by Pettijohn. In this table[7] granitic intrusion with accompanying orogeny is the basis of subdivision, and it is apparent that the interpretations of different workers cannot easily be harmonized.

This is particularly evident when the results of radioactive determinations are used to compare the various granites shown in the table as Algoman. Thus, radioactive evidence indicates that at least part of the Huronian granite in the Sudbury district is much younger than the granites in the Oiseau River area, whereas the granites of southeastern Ontario, classed as Algoman by Miller and Knight, are similarly indicated to be intermediate in age.

[4] H. V. Ellsworth, "Rare-Element Minerals of Canada," *Geol. Surv. Canada Ec. Geol. Ser.*, No. 11 (1932).

[5] J. S. de Lury and H. V. Ellsworth, "Uraninite from the Huron Claim, Winnipeg River Area, Southeast Manitoba," *Amer. Min.*, Vol. XVI (1931), pp. 569–75.

[6] Personal communication.

[7] Prepared in co-operation with H. H. Beach.

EARLY GEOLOGICAL WORK IN THE BIRCH–SLATE LAKES DISTRICT

Dowling,[8] Camsell,[9] and Wilson,[10] in their geological reconnaissance trips through the Birch–Slate Lakes region, at first considered the volcanic and sedimentary schists as Huronian in age and the granites, gneissoid granites, and part of the sedimentary gneisses as Laurentian. Then recognizing (as did Lawson in the Rainy Lake country) that most of their Laurentian in the district was intrusive into their Huronian and that the time relations were not in accord with Logan's application of these terms, they discarded the term "Huronian" in favor of Lawson's "Keewatin," pointing out the lithologic similarity of the rocks in the two regions.

E. M. Burwash,[11] as a result of a traverse through the Birch–Slate Lakes district in 1919, classified the sediments at Birch Lake and Slate Lake as Lower Huronian and for no given reason placed a granite between the "Keewatin" and "Lower Huronian" of the district, which he called Laurentian. In an editorial footnote to the report, Miller suggested that the Lower Huronian of Burwash in this district be referred to the Timiskaming. This was the first time the term "Timiskaming" was extended to the northwestern part of the province, where its use has since been widespread. Two years later Burwash[12] published a new classification for the Birch–Slate Lakes district in which he named the sediments at Birch Lake and Slate Lake the Birch Lake series, correlating them on the basis of lithologic similarity. This correlation, as will be shown, is untenable. At the same time, as pointed out by Pettijohn,[13] after a microscopic ex-

[8] D. B. Dowling, "Report on the Country in the Vicinity of Red Lake and Part of the Basin of Berens River, Keewatin," *Geol. Surv. Canada Ann. Rept.*, Vol. VII, Part F (new ser., 1894).

[9] C. Camsell, "Country around the Headwaters of the Severn River," *Geol. Surv. Canada Ann. Rept.*, Vol. XVI, Part A (1904), pp. 143–52.

[10] A. W. G. Wilson, "Report on a Traverse through the Southern Part of the Northwest Territories from Lac Seul to Cat Lake in 1902," *Geol. Surv. Canada Pub. 1006* (1909).

[11] "A Geological Reconnaissance into Patricia," *Ont. Dept. Mines Ann. Rept.*, Vol. XXI, Part I (1920), pp. 157–92.

[12] "The Pre-Cambrian of Western Patricia," *Jour. Geol.*, Vol. XXX (1922), pp. 393–400.

[13] *Op. cit.*

TABLE 1

CORRELATION CHART SHOWING PRE-CAMBRIAN CLASSIFICATIONS OF VARIOUS AUTHORS TOGETHER WITH THAT ESTABLISHED FOR THE BIRCH–SLATE LAKES DISTRICT

(W. E. Logan, 1863)			Eastern Ontario (W. G. Miller and C. W. Knight, 1914)		Northwestern Ontario (A. C. Lawson, 1913)		Lake Superior Region (C. K. Leith *et al.*, 1935)		
								Killarney granite (?)	
Upper copper-bearing series	Upper group		Keweenawan	Upper copper-bearing rocks of Lake Superior, Upper group (*Geol. Surv. Can.*, 1863) Nipigon, Kaministigua, Nipissing diabase, Sudbury nickel eruptive	Keweenawan (Nipigon)		Keweenawan	Keeweenawan Duluth gabbro Logan sills Osler series Sibley series	
			Unconformity		Unconformity				
	Lower group		Animikean	Upper copper-bearing rocks of Lake Superior, Lower group (*Geol. Surv. Can.*, 1863) Includes Middle and part of Lower Huronian of Lake Huron and various authors (*U.S. Geol. Surv. Bull. 360*) Animikie or Upper Huronian of Lake Superior and Sudbury Basin, Whitewater and Ramsay Lake Series Cobalt Series	Algonkian	Animikie	Algonkian Type	Huronian	Upper: Part of Rove slate Most of Cobalt series of Soo and Batchawwana Middle: Gunflint iron formation Part of Bruce Series Lower: Part of Bruce Series of Soo area
Great unconformity			Great unconformity		Eparchaean interval				
Lower copper-bearing series	Huronian	Huronian granite	(Algoman)	Algoman of Lawson (1913) Laurentian, in part, of some authors and Huronian granite of others. Lorrain granite of Cobalt. Moira granite of Hastings. Killarney granite of Lake Huron	Archaean	Algoman granite		Algoman	Vermillion granite
		Huronian greenstone (basic intrusives) Huronian sediments	Temiskamian	Includes part of Huronian of Logan and Murray and of many other authors Part of Lower Huronian of Van Hise and Leith, (*U.S. Geol. Surv. Bull. 360*) Temiskaming series of Miller Sudbury series of Coleman Matagami, Broadback, Doré, Steeprock, Seine, Wanipigow series, and part of Hastings series		Huronian: Upper / Unconformity / Lower		Knife Lake (may be Lower Huronian)	Knife Lake slate, including most of schist series mapped by Lawson as Coutchiching Windigokan (?) conglomerate and slate. Also Tanton's Coutchiching. Doré conglomerate, Lower part of Bruce series of Soo Area
Great unconformity			Great unconformity		Unconformity				
Laurentian	Upper Laurentian (1865)		(Laurentian) intrusives		Laurentian granite			Laurentian granite	
	Lower (1865)	Grenville sediments Ottawa gneiss	Loganian: Grenville	Includes most of the Hastings and Grenville of older authors, and the Couchiching of Lawson	Ontarian	Keewatin Coutchiching	Archean Type	Keewatin	Keewatin volcanics Ely Greenstone Coutchiching schists
			Loganian: Keewatin	Keewatin complex					

TABLE 1—*Continued*

Sudbury District and North Shore of Lake Huron (W. H. Collins, 1936)			Northwestern Ontario (F. J. Pettijohn, 1937)			Birch–Slate Lakes District (Uchi-Slate Lakes Area, 1939)
Late Pre-Cambrian	Keweenawan	Olivine diabase "Trap" dikes Murray granite Nickel irruptive Conglomerate, sandstone, and lavas	Late Pre-Cambrian	Keweenawan	Osler volcanics Sibley slates	
	Whitewater series					
Great (?) unconformity						
Huronian		Granite: Killarnean(?)				
	Cobalt Series	Upper quartzite Banded cherty quartzite Lorrain quartzite Gowganda formation		Animikie	Rove slate	
	Bruce Series	Serpent quartzite Espanola limestone Bruce conglomerate Mississagi quartzite Ramsay Lake conglomerate			Gunflint iron formation	
Great unconformity			Major unconformity			Major unconformity
Pre-Huronian		Granite (Laurentian, of Coleman; Algoman, of Miller and Knight)			Granitic intrusion	Archean: Younger granite
	Sudbury Series	McKim greywacke	Early Pre-Cambrian	Knife Lake series	Seine Windigokan	
					Major unconformity-granitic intrusion(?)	
				Kenoran volcanics	Neepawa volcanics Handy Lake volcanics Ely greenstone(?)	Archean: Younger volcanics at Shabumeni Lake; basic intrusives in Uchi series
					Disconformity	
				Savant sediments	Abram series Warclub series Crowduck conglomerate Manitou series Steeprock series(?) Birch Lake series(?)	Archean: Younger sediments (Birch Lake series, Timiskaming, Shabu series)
	Interbedding				Major unconformity	Unconformity
	Keewatin	Keewatin, including Copper Cliff arkose [rhyolite]			Granitic intrusion	Confederation granophyre(?)
				Rice Lake Series Lac Seul Series: Keewatin volcanics	Jutten volcanics Beresford Lake volcanics(?) Ely greenstone(?)	Uchi series (volcanic)
					Interbedding	Structural unconformity
				Rice Lake Series Lac Seul Series: Manigotogan sediments	Miniss series Coutchiching	Slate Lake series (sedimentary)
					Unconformity-granitic intrusion	

amination of metasedimentary granulites identified as "soda-rhyolite," Burwash gave the name "Lac Seul series" to part of the Slate Lake metasediments and, on the basis of their supposed volcanic origin, correlated the series with the Keewatin. In this new classification Burwash discarded his "Laurentian" granite and emplaced both an older and younger granite into the sediments and volcanics comprising the schist belt.

The Birch–Slate Lakes district was examined by J. W. Greig[14] in 1926, who in his report refrained from giving any names to the rock series, but in the map accompanying the report the Slate Lake and Birch Lake sediments are indicated as Timiskaming, and the intervening and surrounding volcanics as Keewatin. The present work has established the existence of a thick basal metasedimentary series in the district, parts of which have formerly been called Birch Lake series, Lac Seul series, and Timiskaming, and which is here called the Slate Lake series. Lying with angular unconformity stratigraphically above these sediments are the Uchi volcanics, which previous workers have named Keewatin.

GEOLOGY OF THE UCHI–SLATE LAKES AREA

The table of geologic formations (Table 2) shows the rock units as far as possible in their chronological order, but, owing to the alternation of lithologic types, only the predominating formations or groups of formations are given for the two basement series.

The Slate Lake series, as shown in Figure 2, underlies the southeastern half of the area and consists, for the most part, of metamorphosed impure quartzites and arkoses. These clastic rocks have been altered to mica schists and gneisses near the borders of intrusive granites and to sericite and carbonate schists along zones of shearing. A great lens of granite pebble-bearing conglomerate occurs within the series, marking the accumulation of coarse detritus in response to local uplift. Southeast of Slate Lake a belt of magnetic iron formation provides an additional stratigraphic horizon marker. In the vicinity of Slate Lake three or four belts of pillowed greenstone flows are intercalated with the sedimentary members of the series.

[14] "Woman and Narrow Lakes Area," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXVI (1927), pp. 85–110.

The northwestern half of the map area is underlain by the dominantly igneous rocks of the Uchi series. Basic volcanics, termed

TABLE 2

PRE-CAMBRIAN (ARCHEAN)

Younger intrusives
- Altered basic dikes
- Aplite and silicic dikes
- Quartz-feldspar porphyry, diorite, diorite porphyry
- Pegmatite
- Granite, syenite, quartz monzonite, granodiorite

(Amphibolite)

Intrusive contact—intense folding

Older intrusives
- Basic intrusives—metagabbro and altered peridotite
- Confederation granite porphyry and mylonitized equivalents

Intrusive contact

Uchi series
- Rhyolite and silicic tuff
- Altered andesitic tuff and breccia
- Metabasalt—pillow and amygdaloidal lavas
- Amphibolite and hornblende schists (altered flows)
- Quartzite, graywacke, slate
- Porphyritic andesite tuff and breccia
- Greenstone—hornblende schist, amphibolite and carbonate schist, with local pillow and amygdaloidal structures

Unconformity—gentle folding

Slate Lake series
- Biotite-quartz schist, granulite, paragneiss, staurolite gneiss
- Impure quartzite
- Hornblende gneiss
- Greenstone—pillow lava
- Arkose, slate, sericite schist, carbonate schist
- Conglomerate
- Iron formation
- Impure quartzite, mica schists

greenstones, comprise the greater part of the series and lie interbedded with thin belts of volcanic breccias, tuffs, and flows of porphyritic andesite and rhyolite. An older granite, named the Con-

federation granite porphyry,[15] is found along the western margin of the area, where it occurs as a long, thick, vertical sill. A number of persistent belts of metagabbro, amphibolite, and peridotite occur between the members of the Uchi volcanics as vertical sills.

These sills have been involved in the regional folding of the rock formations. As a result of this same folding the sediments and volcanics now lie in nearly vertical positions. The members of the Slate Lake series strike to the west with a somewhat undulatory trend. The flows and interbedded tuffs of the Uchi series, on the other hand, strike to the south and swing around to the southwest in the southwestern part of the area. Each series is thus a structural unit, the formations of the Slate Lake series abutting against the base of the Uchi volcanics, the contact between them being an unconformity. There is no faulting along the contact.

THE UCHI–SLATE LAKE UNCONFORMITY

The thick sills spaced at intervals within the Uchi volcanics acted as rigid members during folding and, as a result, no fold axes pass through that part of the series within the map area. In consequence, the entire section (20,000 feet thick, excluding the thickness of the sills) has been turned up on edge and is homoclinal; the tops of the beds and flows, as determined from well-developed graded bedding in the tuffs and the shapes of pillows in the flows, consistently face west or northwest. The Slate Lake series is somewhat more complexly folded, but none of the principal formations is repeated, although minor repetitions occur and the tops of the beds for the most part face the north. The pitch of the fold axis is to the west and southwest, and in the minor folds along the unconformity the sedimentary members of the Slate Lake series underlie the volcanics. Complete structural data for these relationships are given elsewhere.[16]

CORRELATIONS WITHIN THE BIRCH–SLATE LAKES AREA

The continuation of the Uchi volcanics northward into the Birch–Springpole Lakes map area was mapped as Keewatin by Harding in

[15] Bateman, "Geology of the Uchi–Slate Lakes Area," *Ont. Dept. Mines Ann. Rept.*, Vol. XLIX (1939).

[16] *Ibid.*

1935;[17] similarly to the northwest, in the Woman and Narrow Lakes map area, Bruce[18] in 1928 referred the volcanics to the Keewatin. In the Birch–Springpole Lakes area the volcanic formations are complexly folded and are overlain by at least one thick infolded sedimentary series, which Harding has called Timiskaming, but to part of which Furse had previously given the name Shabu series and called Timiskaming-type. The writer has discussed the relations of these sediments to the volcanics with Dr. Harding and is satisfied that they are younger. It was pointed out that in the eroded anticlinal structures the volcanics outcropped in the sedimentary areas as through a "window." In the Shabumeni–Birch Lakes area (which constitutes part of the Birch–Springpole Lakes area) Furse[19] found volcanics above and below the Timiskaming-type or Shabu sediments. Furse's "Lower Volcanics" (Harding's Keewatin) correspond to the Uchi volcanics to the south, toward which they strike. His "Upper Volcanics" may be the extrusive equivalents of the basic intrusives that occur to the south, as these intrusives are not reported in the Upper Volcanics.

As a result, the geologic sequence in Table 3 is provisionally established for the Birch–Slate Lakes district. The lithology of the district thus represents an alternating succession of volcanism and sedimentation. In addition, Harding reports the presence of volcanics within the younger (Timiskaming) sediments, and volcanics also comprise a small part of the Slate Lake series. If these minor volcanics within either of the sedimentary series were to be extensively developed in neighboring areas, they would, according to past practice, be called Keewatin. The necessity of restricting the use of Keewatin, if it is to have any meaning other than lithologic, is thus evident.

GEOLOGIC SEQUENCE EAST OF SLATE LAKE

The greater part of the schist belt for 100–150 miles east of Slate Lake has been covered by geologic map sheets, most of which are of

[17] W. D. Harding, "Geology of the Birch–Springpole Lakes Area," *Ont. Dept. Mines Ann. Rept.*, Vol. XLV, Part IV (1936).

[18] E. L. Bruce, "Gold Deposits of Woman, Narrow and Confederation Lakes," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXVII, Part V (1928), pp. 1–51.

[19] G. D. Furse, "Geology of the Shabumeni–Birch Lakes Area," *Ont. Dept. Mines Ann. Rept.*, Vol. XLII, Part VI (1933), pp. 21–51.

a reconnaissance nature. The Papaonga area, lying between Slate Lake and Lake St. Joseph, was examined in 1935 by W. I. Wright, and, although no report has appeared as a result of the work, Wright's map[20] shows quartz-biotite schists indicated stratigraphically below the "Keewatin" volcanics and included under the term "Keewatin." The writer examined these mica schists south of the Uchi–Slate Lakes map area and found them to be part of the Slate Lake metasedimentary series. Otherwise, part of the Slate Lake sediments is shown on Wright's map as Keewatin volcanics. The belt is shown extending almost continuously eastward from Slate Lake to

TABLE 3

ARCHEAN

Younger granite (Algoman of former workers)
Intrusive contact—intense folding
Younger volcanics (Furse); basic intrusives (metagabbro)*
Disconformity
Timiskaming: Shabu series (Furse); Birch Lake series (Burwash)
Disconformity. Confederation Lake granite porphyry†
Uchi series (Keewatin of previous workers)
Unconformity—gentle folding
Slate Lake series (metasedimentary—contains granite-pebble-bearing conglomerate, iron formation, and some volcanics)

* The basic intrusives in this district have been formerly designated Keewatin, Algoman, pre-Algoman, and post-Keewatin, depending on what other formations were examined in a particular area.

† The position of the Confederation Lake granite porphyry is tentative, as the relation of this intrusive to the younger sediments is not known.

Lake St. Joseph and, throughout the greater part of this distance there are indicated pre-volcanic metasediments lying to the south of the "Keewatin."

At Lake St. Joseph, Bruce studied the paragneisses and included iron formations and granite-pebble conglomerate,[21] which he showed were stratigraphically below the volcanics lying to the north. Bruce gave no names to the formations, but in compiling the first edition of the Kenora sheet Tanton,[22] assuming the volcanics are Keewatin, classed the lower sediments as Coutchiching. In the Miniss-Pash-

[20] "Papaonga Area," *Geol. Surv. Canada Map 347A* (1936).

[21] "Iron Formation at Lake St. Joseph," *Ont. Dept. Mines*, Vol. XXXI, Part VIII (1922), pp. 1–32.

[22] T. L. Tanton, "Kenora Sheet, Ontario," *Geol. Surv. Canada Map 266A* (1933).

kokogan area, which includes part of Lake St. Joseph and the district to the south, Dyer[23] concluded that the metasediments in the southern part of the area are pre-volcanic. Dyer called the volcanics Keewatin and gave the name "Miniss series" to the older sediments, which he considered the equivalent of the Coutchiching. Thus, for 150 miles east of Slate Lake there is indicated a pre-volcanic metasedimentary series along the southern part of the schist belt and on the strike of the Slate Lake series.

GEOLOGIC SEQUENCE WEST OF SLATE LAKE

The Slate Lake metasediments to the west are continuous with similar rocks that have been mapped by Bruce[24] in the English River Valley and examined by Burwash at the head of Lac Seul. To the north at Red Lake, Bruce[25] studied a volcanic series (Keewatin) overlain by infolded younger sediments (Timiskaming). Horwood, who has recently studied the area in detail, has not yet published but has shown to the writer's satisfaction that the sediments at Red Lake are postvolcanic. Although Bruce did not give a name to the Slate Lake equivalents in the English River Valley, he later indicated[26] that they might be part of a great pre-Keewatin deposit—the Coutchiching delta—and still later[27] suggested a pre-volcanic age for these sedimentary gneisses. This suggests that from the upper English River to Red Lake there is an alternation of volcanics and sediments resembling that in the Birch–Slate Lakes district, although no equivalents of the Upper Volcanics at Shabumeni Lake have been reported from Red Lake.

In the Minaki–Sydney Lakes area, which covers the lower English River Valley to the Manitoba boundary, D. Derry[28] mapped a

[23] W. S. Dyer, "Geology of the Pashkokogan–Misehkow Area," *Ont. Dept. Mines Ann. Rept.*, Vol. XLII, Part VI (1933), pp. 1–20.

[24] "Geology of the Upper Part of the English River Valley," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXIII, Part VIII (1924), pp. 1–11.

[25] E. L. Bruce and J. E. Hawley, "Geology of the Basin of Red Lake," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXVI, Part III (1927; reprinted 1932), pp. 1–72.

[26] "Coutchiching Delta," *Bull. Geol. Soc. Amer.*, Vol. XXXVIII (1927), pp. 771–82.

[27] "Geology of the Red Lake and Woman Lake Gold Areas, Northwestern Ontario," *Amer. Inst. Min. Met. Eng. Tech. Pub. 107* (1928).

[28] "Geology of the Area from Minaki to Sydney Lake," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXIX, Part III (1930), pp. 22–41.

large area of metasediments and volcanics and, although he did not establish the age relationships, concluded that the sedimentary rocks are in part older and in part younger than the volcanics. Previous to this, Gilbert,[29] working to the north in the Gammon River area on the extension of the Rice Lake series into Ontario, gave the name "Keewatin-Timiskaming" to the volcanic-sedimentary complex. Gilbert stated that the relationships between the different formations were not known but would probably be clarified by the work J. F. Wright was doing at that time on the Manitoba side of the boundary. To the south Derry grouped all the rocks of the schist belt, both volcanic and sedimentary, as "Keewatin(?)."

Burwash[30] in a survey of part of the Ontario-Manitoba boundary used the classification previously adopted by Moore for southeastern Manitoba. Moore[31] originally mapped a large reconnaissance area in southeastern Manitoba, separating the basement rocks into the Rice Lake series (volcanic) and the Wanipigow series (sedimentary), which was said to be younger and to occupy a syncline in the Rice Lake series, but no reliable evidence was offered to support this conclusion. Dresser[32] later stated that Moore's conclusion regarding a younger age for the Wanipigow series was based on evidence obtained from outside the Rice Lake area proper and that it was not yet clear which of the two series is the older. In 1921 Cooke[33] studied two small areas—one at Rice Lake and the other at Oiseau River. He stated that within his map areas the Wanipigow sediments are bedded tuffs which rest with conformity on the surface flows[34] and suggested that at Rice Lake the sediments are both underlain and

[29] G. Gilbert, "Gammon River Area and Rickaby Lake Schist Belt," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXVI, Part III (1927), pp. 73–84.

[30] "Geology of the Ontario–Manitoba Boundary, Winnipeg River to Bloodvein River," *Ont. Dept. Mines Ann. Rept.*, Vol. XXXII (1923), pp. 1–47.

[31] E. S. Moore, "Region East of the South End of Lake Winnipeg," *Geol. Surv. Canada Summ. Rept.* (1912), pp. 262–70.

[32] J. A. Dresser, "Gold-bearing District of Southeastern Manitoba," *Geol. Surv. Canada Summ. Rept.* (1916).

[33] "Geology and Mineral Resources of the Rice Lake and Oiseau River Areas," *Geol. Surv. Canada Summ. Rept.*, Part C (1921), pp. 1–36.

[34] Stockwell has recently indicated that these bedded rocks at Rice Lake are post-intrusive feldspathic quartzites that lie with angular unconformity above the lavas—the San Antonio formation (H. C. Stockwell, "Rice Lake Gold Area," *Geol. Surv. Canada Mem. 210* [1938]).

overlain by volcanics; consequently, he used the term "Rice Lake series" in connection with all the rocks of the schist belt.[35] From 1922 to 1929 J. F. Wright did considerable field work in southeastern Manitoba, and his report[36] contains the most complete information available on the stratigraphy. Using reliable criteria for the determination of the tops of beds and flows, Wright interpreted the Beresford–Rice Lake area as the south limb of a broad syncline, and the Oiseau River area as the broad north limb, the intervening granite containing long, narrow schist inclusions—suggesting intrusion along a broad and upfolded arch of the sediments. As a result, Wright classed all the pregranite volcanic and sedimentary formations as the Rice Lake series, which he subdivided into a lower Manigotagan[37] sedimentary phase, an intermediate Beresford volcanic phase, and an uppermost Wanipigow sedimentary phase.

Thus, for 150 miles west of Slate Lake, the rocks along the southern part of the schist belt are predominantly sedimentary and, where the age relationships have been proven, are also pre-volcanic. Pettijohn has reached essentially the same conclusion, except in the case of the Slate Lake sediments, which he accepted as belonging to a younger group. He states:

> The sediments, largely paragneisses, which are found at intervals from the Oiseau River area in Manitoba, through the valley of the English River, on the western end of Lac Seul and north of Lac Seul, and on the south shore of Lake St. Joseph, and eastward, are probably one and the same series. They are along the same general strike line, and some continuity exists between the several known outcrops, which extend for about 300 miles. From all the evidence available, it seems probable that they lie below the associated volcanic rocks and their schist equivalents.[38]

[35] Owing to the small size of the areas studied by Cooke, his conclusions are not necessarily applicable to the district as a whole.

[36] "Rice Lake Map Area, Southeastern Manitoba," *Geol. Surv. Canada Summ. Rept.*, Part C (1922), pp. 45–82; "Geology and Mineral Prospects of the Northern Part of the Beresford Lake Map Area, Southeast Manitoba," *Geol. Surv. Canada Summ. Rept.*, Part B (1923), pp. 86–104; "Geology and Mineral Deposits of Oiseau River Map Area, Manitoba," *Geol. Surv. Canada Summ. Rept.*, Part B (1926), pp. 51–104; "Gold, Copper-Nickel and Tin Deposits of Southeastern Manitoba," *Geol. Surv. Canada Summ. Rept.*, Part B (1929), pp. 136–71; "Geology and Mineral Deposits of a Part of Southeastern Manitoba," *Geol. Surv. Canada Mem. 169* (1932).

[37] Moore originally applied the term "Manigotagan" to the granitic rocks of the district.

[38] *Op. cit.*

TABLE 4

TENTATIVE CORRELATIONS OF THE DIFFERENT VOLCANIC AND SEDIMENTARY SERIES ALONG THE RICE LAKE–LAKE ST. JOSEPH SCHIST BELT

West Rice Lake and Oiseau River Area		Lower English River Valley	Red Lake and Upper English River Valley	Birch–Slate Lakes District	Papaoga Area	Lake St. Joseph Area	*East* Pashkokogan-Misehkow Area
							Keweenawan
Intrusives		Algoman (?)	Post-Timiskaming intrusives (Algoman)	Postvolcanic intrusives: Younger granite	Granite	Granite	Algoman
Rice Lake Series	Wanipigow sediments*	Timiskaming or Keewatin–Timiskaming complex; some sediments are reported pre-volcanic and others postvolcanic but relations have not, as yet, been established	Timiskaming	*Intrusive contact* Basic intrusives Upper volcanics *Disconformity* Younger sediments†	Sediments		Timiskaming
			Postvolcanic diorite	Older granite (?) *Intrusive contact*			
	Beresford volcanics		Keewatin	Uchi volcanics *Unconformity*	Keewatin	Volcanic group	Keewatin
	Manigotagan sediments		Metamorphosed sedimentary schists and gneisses, iron formation and conglomerate of upper English River Valley	Slate Lake sediments	Quartz-biotite schist	Sedimentary group of paragneisses	Miniss series (Coutchiching) consisting of quartz-biotite gneiss

* Part of Wanipigow sedimentary phase has since been indicated as postintrusive (see Stockwell, *Geol. Surv. Canada Mem. 210* [1938]).

† Younger sediments variously called Birch Lake series, Shabu series, or Timiskaming.

The writer is in complete agreement with this statement and can only add that the conclusions are strengthened by the evidence of the Slate Lake series being pre-volcanic in age. It is also probable that a large part of the volcanics overlying this series are approximately contemporaneous in age. The younger sediments, however, such as the Wanipigow series, the sediments at Red Lake, and the younger sediments at Birch Lake, are largely in the form of discontinuous lenses and may be of somewhat different ages. Except for the San Antonio formation, all these rocks have been involved in widespread folding and granitic injection. The only prefolding granitic rock reported along the belt is the Confederation Lake granite porphyry. The geologic succession as shown in Table 4 for the different areas shows a preliminary attempt at correlating the major series along the belt.

CONCLUSIONS

The volcanic-sedimentary relationships that have been established at Rice Lake, Slate Lake, and Lake St. Joseph indicate the existence of a great pre-volcanic sedimentary series, as it seems probable that this relationship exists in the intervening areas along the strike of the formations, suggesting that the series is continuous.

The term "Timiskaming," first applied by W. G. Miller[39] to certain ancient pre-Cambrian sediments in northeastern Ontario, has been widely used by Canadian geologists.

Subsequent geological work has shown that part of the Timiskaming is homoclinal in the "Keewatin," part is separated from the "Keewatin" along an angular unconformity, but most of it is younger than the "Keewatin" and separated from the underlying volcanics either by a basal conglomerate or by a gradational zone representing the passage of a period of volcanism into one of sedimentation. Recent work by H. C. Gunning and J. W. Ambrose[40] in the Rouyn-Harricanaw area of Quebec has shown that the bulk of the "Timiskaming" sediments in that area underlie the "Keewatin" lavas and that the early pre-Cambrian rocks are represented by an alternating succession of volcanics and sediments in conformable arrangement.

[39] "Notes on the Cobalt Area," *Eng. and Min. Jour.*, Vol. XCII (1911), pp. 647–48.

[40] *Trans. Roy. Soc. Canada*, Vol. XXXIII, Sec. IV (1939), pp. 19–49.

The lithology of the Birch–Slate Lakes district represents an alternation of sedimentation with volcanism and does not conform to the simple Keewatin-Timiskaming concept of an earlier volcanic period followed by a period of sedimentation. As the younger sediments, such as those at Red and Birch lakes, may lie at different horizons within the volcanics, they are not necessarily of the same age; thus Timiskaming cannot be applied as a strictly time term to these sediments.

The name "Keewatin" has been used extensively in connection with the volcanics along the Rice Lake–Lake St. Joseph belt, including the Uchi series. While a correlation of these volcanics with those in the type Keewatin area cannot be proven, owing to the intervening granite, most workers assume an approximate equivalence in age. Since Lawson defined the Coutchiching as a pre-Keewatin sedimentary series in 1887, much of the series has been shown to be post-Keewatin, and there has been much dispute as to whether or not there are any pre-Keewatin sediments in the type area of the Coutchiching. The pre-volcanic age of the Manigotagan–Slate Lake–Miniss series is not disputed. Consequently, according to Lawson's definition, if anyone applies the term "Keewatin" to the principal volcanic areas along this belt, then the pre-volcanic sediments, such as the Slate Lake series, should be considered Coutchiching.

ARTIFICIAL HELICTITES AND GYPSUM FLOWERS

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ABSTRACT

Deposits quite similar to natural helictites were formed in the laboratory by allowing solutions of different salts to evaporate. These deposits indicate that natural helictites grow at the free end and assume their crooked shape as a result of the chance orientation of crystals. Artificial formations similar to gypsum flowers suggest that the natural gypsum flowers grow at the attached end and assume their curvature, not as a result of strain, but by a more rapid growth on one side than on the other.

INTRODUCTION

The word helictite comes from a Greek root meaning "a spiral." It is applied to small deposits on cave walls and ceilings that are not straight like stalagmites and stalactites but bend irregularly in different directions (Figs. 1 and 2). Gypsum flowers, also occurring on cave walls and ceilings, are composed of bundles of acicular gypsum fibers that curve outward from a common center (Figs. 7 and 8).

Even though helictites and gypsum flowers are not so common as many other types of cavern deposit, their peculiar and interesting shapes drew the attention of the earliest cavern explorers. There has been much theorizing on their formation, but, as yet, little uniformity of opinion has been reached. Even the fundamental question of whether helictites grow at the base or at the free end has not been settled. So elementary a problem ought to be solved easily by experimentation, even though the natural conditions cannot be reproduced exactly.

Various stages of development can be seen in caves, but growth proceeds very slowly, and interpretation is difficult. If the process can be reproduced artificially, however, rapidly enough for continuous development to be observed, the interpretation will be considerably facilitated. The experimental method is also valuable because the conditions can easily be varied and the effect noted.

The low solubility of gypsum and calcite practically forbids their use in experimental approach to the problem. This difficulty can be

FIG. 1.—Group of helictites in Wyandotte Cave, Indiana

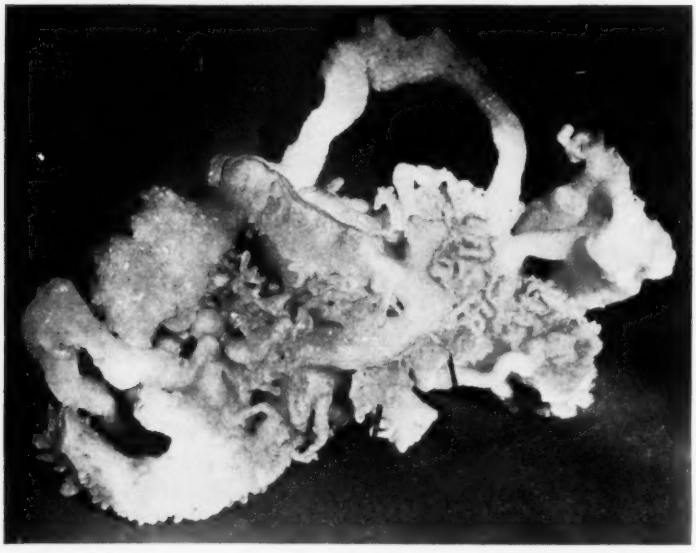


FIG. 2.—Helictites from Copper Queen Mine, Bisbee, Arizona. This formation is 11 inches long.

avoided by the use of more soluble salts. If it is found that a large number of soluble salts all act the same way, then it may be inferred, with reservations, that the less soluble ones will act similarly. A much stronger basis for assuming similarity of process of the natural and of the artificial formations would be a similarity in the end products. It would be a remarkable coincidence if close similarities in form could be produced by entirely different processes.

ARTIFICIAL HELICTITES

In order to grow artificial helictites the solution must be supplied so slowly that it will not drip from the formation. Several methods for obtaining this small flow were tried. The method that worked best consisted of placing a cork fairly loosely in the bottom end of a vertical open tube, into which the solution was then poured (Fig. 3). The cork was porous enough to allow the solution to seep slowly out of the end of the tube. Upon reaching the surface, the water evaporated, and its dissolved salt was precipitated.

When the flow of solution was fast enough it collected as a drop on the end of the cork. Any salt deposited formed a ring around the base of the drop. This process, continued, built the ring into a cylinder, and a deposit very much like a small "straw" stalactite was formed.

When, however, the flow was slow enough, evaporation prevented the formation of a drop, and the deposit was not oriented by gravity. Deposition took place around the point of emergence of solution. It formed a ring, but one smaller than that formed around the pendant drop. This ring also grew into a tube. The slow flow and consequent slow evaporation allowed the growth of crystals, as the tube wall increased, that were large relative to the size of the opening. When a crystal happened to grow transversely to the length of the tube, the tube suffered a change of direction of growth.

When a crystal grew so tightly across the end of the tube that it impeded the flow, hydrostatic pressure in the tube above built up until the solution broke through the side of the tube somewhere to produce a branch. A continuation of this process produced medusa-like forms resembling those in many caves.

The artificial formations looked very much like the natural

helictites. They were tubes ranging from 2 to 5 mm. in diameter, whose crooked shape showed that their orientation had been little

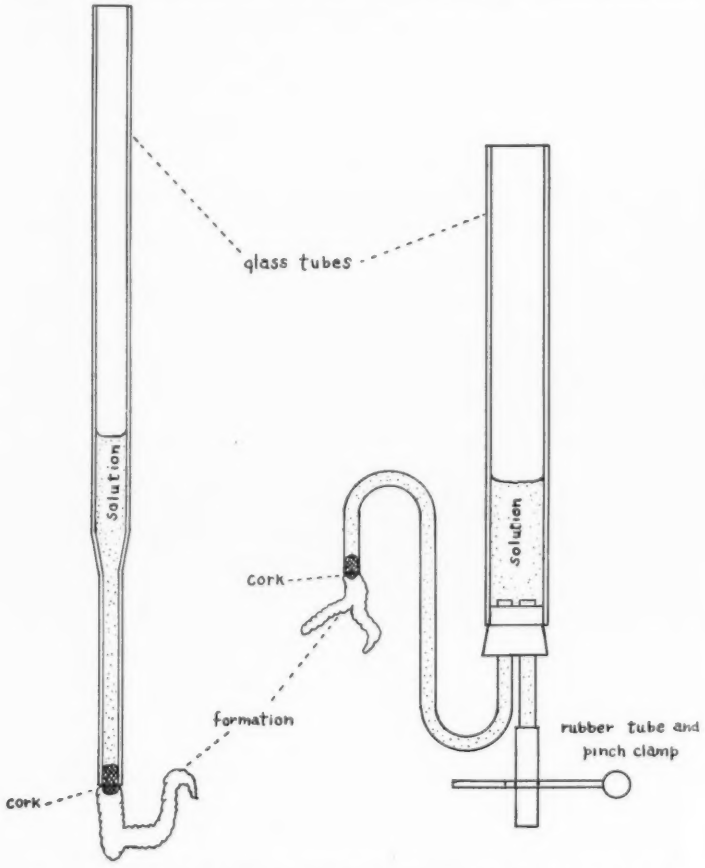
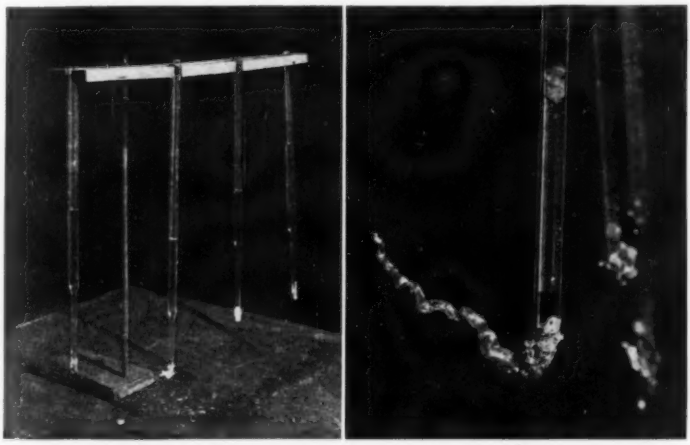


FIG. 3.—*Left:* simple apparatus for forming artificial helictites. *Right:* helictite apparatus modified so that the hydrostatic head can be easily controlled.

affected by gravity. They branched in a complicated manner as helictites do. Natural formations that extend from the cave wall around an irregular loop and then back to the cave wall are often

seen. There were analogous cases among the artificial formations. Many returned to the glass after describing an irregular arc. Some crept up the side of the glass, utilizing the glass itself as one side of the helictite tube, and then left the glass for some unknown reason.

Evidently, there is a complete transition from helictites to stalactites. The nature of the deposit depends upon the rate of flow of the solution; when the flow is large a stalactite is produced, and when it is small a helictite is produced. Often, in nature, helictites



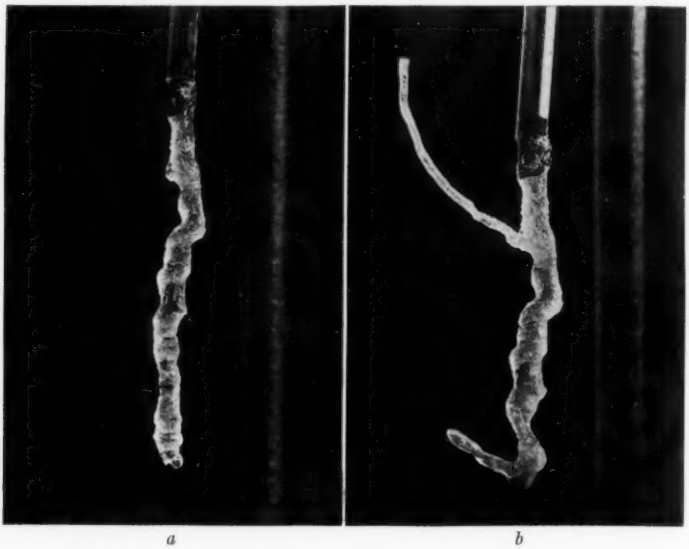
a *b*

FIG. 4.—(*a*) Apparatus for growing artificial helictites. The glass tubes are 15 inches long. (*b*) Artificial helictite composed of potassium dichromate. The glass tube at the top has an outside diameter of $\frac{1}{4}$ inch.

grow from the sides of stalactites, indicating a reduction of supply of solution. The reverse was not found, presumably because the channel can readily diminish in size but it can only rarely increase in size.

This relation is shown by the artificial sodium thiosulphate formation in Figures 5*a* and 5*b*. When it started growing, the flow was sufficient to produce a straight, stalactite-like tube. After this had grown several inches, a valve controlling the supply of solution was almost completely closed. The rate of growth immediately decreased, and the later growth was of the nature of a helictite rather

than of a stalactite. The same relation was illustrated by the artificial formations in another way. When a number of experiments were started simultaneously, the ones that developed the largest deposits were like stalactites, and the others were like helictites. The amount of deposit roughly indicated, of course, the rate of flow of the solution.



a *b*

FIG. 5.—(*a*) This nearly straight formation corresponds to the small "straw" stalactites. The glass tube at the top has an outside diameter of $\frac{1}{4}$ inch. (*b*) This is the same tube plus helictite-like growths that were formed when the flow of solution was decreased by partially closing a valve.

The artificial helictites can be grown with many different salts. Sodium thiosulphate, potassium dichromate, calcium chloride, and copper sulphate were all tried successfully in the laboratory. Saturated solutions produce the most rapid growth, but other concentrations can be used. The only marked difference between these artificial helictites and the natural calcium carbonate ones is the difference in growth rates.

The literature reveals a difference in opinion as to whether helictites grow at the free end or at the attached end. One of the early

students of caves—John Collett (1878)[1]—described them as growing by propulsion from the bottom. G. P. Merrill (1894)[2] and W. S. Blatchley (1896)[3] took the opposite view. R. J. Holden (1938)[4] agreed with Collett.

If the process which forms artificial helictites is the same as that which produces the natural ones, then the latter must also grow at the free end. When the artificial ones are forming it is very easy to observe the manner of their growth from day to day. In every case, unless a branch was forming, the entire growth took place at the free end.

There are additional reasons for believing that this should be the case in the natural formations. When it is assumed that the entire growth of the formation takes place at the base, it is also assumed that there is no travel of solution along the stem of the helictite, because if there were the water would undoubtedly evaporate and deposit its load of salt. Helectites that branch are very common. If branches originate by growth at the base, it must be postulated that two sprigs started growing at two adjacent points and migrated toward each other as growth continued until they joined, which is very unlikely.

The peculiar twisted form of helictites has been explained in a number of different ways. H. C Hovey (1882)[5] attributed it to lateral outgrowths having fungi for a starting-point. C. S. Dolley (1886)[6] advanced the idea that the deposition took place in a drop of water held by the filament formed by a small cave spider. Every bend in the formation would thus indicate the breaking of the old filament and the formation of a new one with a different orientation. Merrill (1894) gave this extensive description:

> They occur sometimes singly, but more commonly in groups, or clusters of several, ranging in sizes from 3 to 10 mm. in diameter. Closer inspection

[1] "Wyandotte Cave," *Ind. Geol. Surv., 10th Ann. Rept.* (1878), pp. 475–76.

[2] "On the Formation of Stalactites and Gypsum Incrustations in Caves," *Proc. U.S. Nat. Mus.*, Vol. XVII (1894), pp. 77–81.

[3] "Indiana Caves and Their Fauna," *Ind. Geol. Surv., 21st Ann. Rept.* (1896), pp. 160–61.

[4] "Helictites in Virginia" (abst.), *Bull. Geol. Soc. Amer.*, Vol. XLIX, No. 10 (1938), p. 1886.

[5] *Celebrated American Caverns* (Cincinnati: R. Clarke & Co., 1882), p. 165.

[6] "On the Helictites of Luray Cave, Virginia," *Proc. Acad. Nat. Sci., Phila.* (1886), pp. 351–52.

reveals the fact that while in most cases tubular, the tube itself is of almost microscopic proportions, being as a rule less than half a millimeter in diameter. So small is it, in fact, that capillarity, not gravity, is the controlling principle in giving direction to the lime-carrying solution. A small spicule of calcite crystallizing on the extremity is as likely to point any other direction as downward: the direction of the next drop is controlled in part by the first, where the same process is repeated. Or on the assumption that the stalactite increases in length by constant additions to the tube, on all sides, it is easy to imagine that the deposit takes place, for a time more rapidly on one side than on the other, perhaps partially closing the orifice or giving it a different direction. The essential fact is, however, that it is to capillarity, and not to gravity, that is due the peculiar vermicular forms. Why, at the outset, the stalactite should begin to form through many small capillary tubes rather than through one larger, as is ordinarily the case, I will not pretend to say. It is to be noted, however, that in Wyandotte, the roof forming limestones are nearly horizontal, while in Luray and many other caves they are highly tilted. This results in a more even percolation of the water in the first instance, the roof being more homogeneous. It is possible, therefore, that the water gathers in drops of smaller size, and very likely in smaller amounts. I have no other than hypothetical data for this last assumption, however.[7]

Blatchley (1896)[8] explained the twists of the formation by the varying currents of air. These changing air currents, according to him, force the tiny drops of water on the tip of the formation first to one side and then to the other and thus determine the direction of growth. F. C. Greene (1906)[9] suggested that the curves are possibly due to the twinning tendency in the crystallization of the calcite. A. K. Lobeck (1929)[10] and W. M. McGill (1933)[11] ascribe the formation of helictites to impurities. Lobeck says:

> Branching forms, known as helictites, result probably from the presence of impurities in the dripping water. These impurities, such as grains of sand or silt, accumulate on the outside of the stalactite and cause a change in the direction of crystal growth at those points.[12]

These different opinions indicate that there are nearly as many

[7] *Op. cit.*, p. 80. [8] *Op. cit.*

[9] "Caves and Cave Formations of the Mitchell Limestone," *Proc. Ind. Acad. Sci.* (1908), p. 182.

[10] "The Geology and Physiography of the Mammoth Cave National Park," *Ky. Geol. Surv.*, Vol. XXXI (6th ser., 1929), p. 380.

[11] "Caverns of Virginia," *Va. Geol. Surv. Bull. 35* ("Educational Ser.," No. 1, 1933), p. 121.

[12] *Op. cit.*

explanations as there are observers. If we assume that the same process operates in the formation of the natural helictites as in the artificial ones, we should be able to test the different explanations by seeing whether or not they can apply to the artificial formations. If we do this we can discard the explanations using fungi and cave spiders, because neither were present in the laboratory. As all the laboratory windows were kept closed we may also discard the theory of varying air currents. The twinning tendency of calcite cannot be used as an explanation because no calcite was used in the formation of the artificial helictites. There were no grains of sand or silt in the laboratory solutions, so the theory based upon the presence of impurities does not apply. There remains just one explanation—that of Merrill—that can apply to both. We then might conclude with Merrill that the chance orientation of the crystals at the end of the tube will determine the direction of its growth. Some of the other conclusions reached by Merrill can be investigated further in the light of the artificial formations.

Merrill's explanation for the growth of helictites is based on the presence of a tube running through the formation. The presence of a tube was also noticed by Dolley (1886).[13] The artificial formations have a large, well-developed tube, through which all the solution passes. Some of the natural helictites clearly show a tube, but others do not seem to show one at all. It is possible, in the latter case, that the original tube has been obliterated by recrystallization of the calcium carbonate. This would not necessarily mean that all flow would stop; it might continue in the numerous crevices between the various crystals.

The occurrence of internal flow in natural helictites that show no obvious tube can sometimes be established. Most of such cases, as in tube helictites, have cross sections with a concentric structure near the periphery. These concentric deposits were evidently formed after the central core. The depositing solution could have come either down the outer surface of the formation or else along the inside and reached the outside by slowly permeating through the crystal interspaces of the side wall of the helictite. If the former possibility were true, the deposits would not be concentric on hori-

[13] *Op. cit.*, pp. 351–52.

zontal stems because gravity would make them thicker on the bottom.

Another type of evidence leads to the same conclusion. If a helictite is found growing from the side of a stalactite, the stalactite will almost always be found to be one of the small, cylindrical, "straw" stalactites. "Straw" stalactites are the only type in which all the solution comes down the central tube. This suggests that the solution is not supplied to the outside surface of the helictite but to the inside. The helictite is merely a crooked branch or extension of the tube.

In his description of the formation of helictites Merrill (1894)[14] stated that capillarity is the controlling principle in giving direction to the lime-carrying solution. O. C. Farrington (1901)[15] also emphasized the importance of capillarity. It is possible to calculate just how far capillarity is able to lift a solution.

$$h = \frac{2\gamma}{rg\rho}$$

$$h = \frac{2 \times 73.5}{0.025 \times 1 \times 980}$$

$$h = 6.0 \text{ cm.}$$

h = capillary rise in cm.
r = radius of tube in cm.
ρ = density of the liquid in gms/cm^3
g = acceleration due to gravity in cm/sec^2
γ = surface tension in dynes/cm

For pure water at a temperature of 15° C. and a tube diameter of $\frac{1}{2}$ millimeter, the capillary rise will be about 6 centimeters. The presence of a solute alters the rise only slightly. Both the artificial formations and the natural ones, however, often have upward bends that are longer than 6 centimeters. Some of the artificial ones were almost twice this length.

In an effort to produce artificial helictites in which capillarity was the major force in moving the solution through the formation, various attempts were made to grow the formations with a very small hydrostatic head. Every time this was done, the salt would deposit as a crust over the solution, and, since there was not enough

[14] *Op. cit.*, p. 80.

[15] "Observations on Indiana Caves," *Field Columbian Mus Geol. Ser.*, Vol. I, No. 8 (1901), p. 251.

pressure to break through the crust, the growth would cease.[16] Although capillarity is undoubtedly important in the growth of helictites, it is clear that a hydrostatic head is more important.

We have seen that a very small flow is required to grow helictites, which means that there must be a very small opening. It may be that there is a fundamental relationship between the small opening and the hydrostatic head. The small hole will become plugged tightly with deposit unless a pressure is placed on the solution by means of a hydrostatic head. The pressure may prevent the tube from plugging because the pressure will tend to break through any barrier of the salt as soon as it is formed. An increase in the hydrostatic head will, of course, mean that the tube size must be decreased to produce the same small flow.

The temperature and relative humidity of caves are practically constant. It is, therefore, rather interesting to find that both calcite and aragonite helictites are known. Dolley (1886)[17] described some helictites from Luray Cavern as aragonite but gave no reasons for his belief. Farrington (1901)[18] described some Wyandotte Cave helictites as calcite on the basis of specific gravity. The writer has in his possession helictites of both kinds (Fig. 6). The calcite ones are from Luray Cavern, Virginia, and those of aragonite are from Floyd Collin's Crystal Cave, Kentucky. They were distinguished on the basis of the reaction with ferrous sulphate, cobalt nitrate (Meigen's reaction), specific gravity, and optical characteristics. The specific gravity was rather low for each type, probably because of considerable pore space. It averaged 2.75 for the aragonite and 2.52 for the calcite. E. S. Dana[19] gives values of 2.94 for the aragonite and 2.71 for the calcite.

According to Dana aragonite will form when solutions of calcium carbonate are evaporated at temperatures of from 80° to 100° C, but it will form at lower temperature if the solution contains some sul-

[16] Helictites are very well developed in Floyd Collin's Crystal Cave, Kentucky. They grow from a travertine crust that covers the roof. One of the guides of this cave stated that in opening a passageway through the cave it was necessary to cut through a portion of this crust and that the first hole made through the crust released a considerable quantity of water.

[17] *Op. cit.*, pp. 251–52. [18] O. C. Farrington, *op. cit.*, pp. 249–50.

[19] *A Textbook of Mineralogy* (2d ed. rev. W. E. Ford; New York: John Wiley & Sons, 1932), pp. 512–21.

phate. Johnston, Merwin, and Williamson obtained aragonite in the presence of sulphate at temperatures as low as 19° C. They explain this by saying that a solid solution of $CaSO_4$ in aragonite may reduce its solubility and increase its stability with respect to calcite.[20] The aragonite helictites were tested for sulphate with barium chloride, and it was found to be present. Perhaps this explains the formation

FIG. 6.—*Left:* calcite helictite from Luray Cavern, Virginia. *Right:* aragonite helictite from Floyd Collins' Crystal Cave, Kentucky. The helictite on the left is 2½ inches long.

of aragonite helictites at cave temperature. The ground water near Floyd Collin's Crystal Cave must have a high sulphate content because much gypsum is deposited in the cave.

ARTIFICIAL GYPSUM FLOWERS

Artificial deposits that correspond to gypsum flowers can be grown by allowing a solution of sodium thiosulphate or other salt to permeate slowly a porous plate and evaporate on its surface. The rate of deposition will, of course, depend upon the rate of flow of

[20] John Johnston, H. E. Merwin, and E. D. Williamson, "The Several Forms of Calcium Carbonate," *Amer. Jour. Sci.*, Vol. XLI (4th ser., 1916), p. 482.

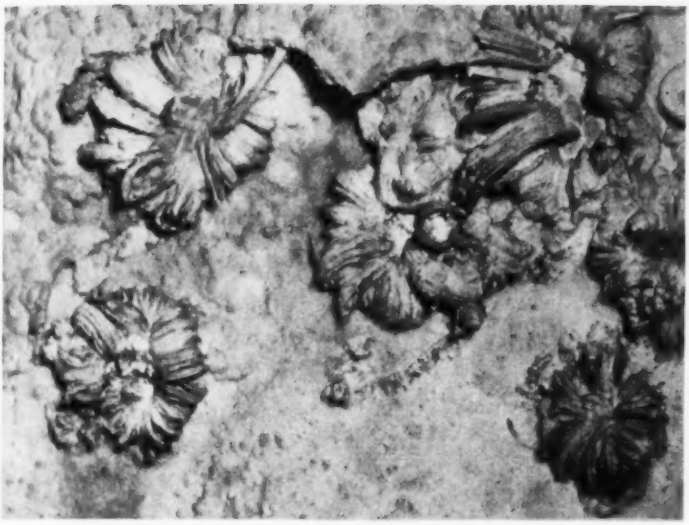


FIG. 7.—Gypsum flowers near the "Snowball Dining-Room," Mammoth Cave, Kentucky. These formations average about 6 inches in diameter.

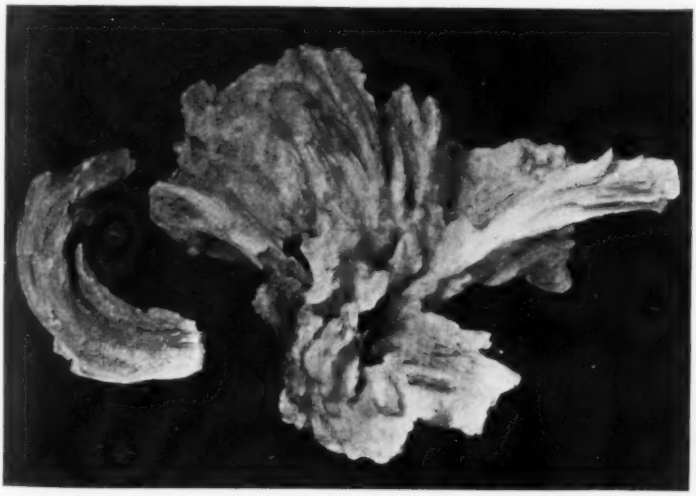


FIG. 8.—Gypsum flower and single petal from Mammoth Cave, Kentucky. The gypsum flower measures 5 inches across.

the solution and the rate of evaporation. The flow must be slow enough, and the evaporation fast enough, for the deposition to take places in the pores of the plate. If the flow is too fast the solution will form a film covering the entire surface of the plate, and large crystals will be obtained instead of the desired formations.

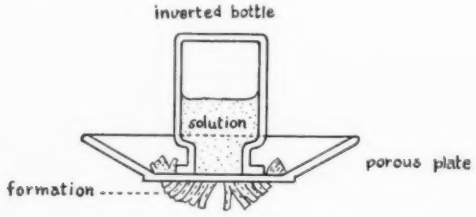


FIG. 9*a*.—Simple apparatus for growing formations like gypsum flowers

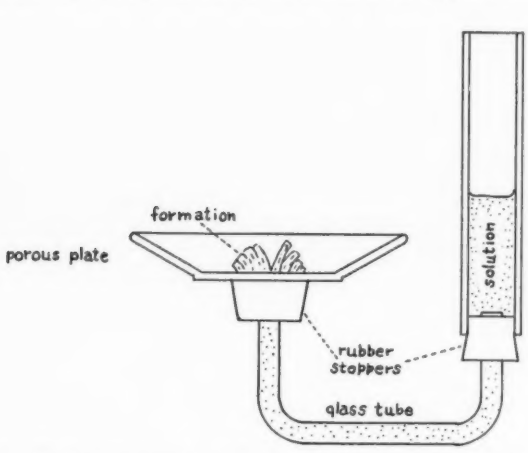


FIG. 9*b*.—Similar apparatus in which the hydrostatic head can be controlled

The porous plates employed are of the sort commonly used in chemical laboratories for drying organic precipitates. Since the flow must be slow and since the porous medium offers innumerable passageways for the solution, the hydrostatic head must be small. The small hydrostatic head was obtained according to the two methods illustrated by Figures 9*a* and 9*b*. In the original method the solution

was placed in a small bottle which was inverted on the plate. The modified method consisted of cementing a rubber stopper to the bottom of the plate with rubber cement and feeding the solution up through the hole in the rubber stopper by means of a glass tube. The glass tube connected with a reservoir in which the level of the solution could be varied at will.

FIG. 10.—Artificial "gypsum flowers" composed of sodium thiosulphate. The apparatus has been inverted for ease in photographing. The diameter of the unglazed porcelain plate is $4\frac{3}{4}$ inches.

When the evaporation took place at the proper rate, continuous deposition occurred in the pores, and the deposit was forced out of the pores by new deposit forming below it. As growth continued, the deposit elongated into needle-like crystals, always growing at their base. Each one bore a resemblance to tooth paste being forced from a tube. The deposit first appeared on the porous plate as a thin, smooth, pearly surface. As growth continued, irregularities in the rate of growth became more apparent, and the surface became undulating. Finally, a break appeared on the surface, and the curved

nature of the crystals could be seen. The breaks always occurred where the deposits were the thickest.

If the pores in which the crystal needles were growing were not very close together, the needles remained separate, and the aggregation had a "furry" appearance. If, however, the pores were close together the needles remained in contact with each other, and the aggregate looked very much like the satin spar of natural gypsum flowers. When the closely packed crystals were short they were perpendicular to the porous plate, but when they grew longer they split and curved like the petals of a gypsum flower. The crystals did not grow out straight for a certain distance and then curve; they had a constant curvature throughout, which became much more apparent as the crystals lengthened. In some cases a small group of pores produced a single, curved fiber. A natural homology of this would be gypsum flowers like the "tiger lily" in Mammoth Cave.

The fact that gypsum rosettes grow at their base and not at their tip has long been recognized. On the cause of their curvature, however, there has not been a uniformity of opinion. Hovey (1882) said that this curvature was ". . . . produced, it is probable, by the crowding of fibrous gypsum, that abounds in these dry halls, through crevices too small for the crystals to grow in; their efforts to spread apart starting curves that are continued by larger number of atoms being supplied to one side than to the other."[21]

Merrill (1894) described their growth as follows:

> The method of growth of these forms is plainly by additions to the bottom, or more properly, to the end attached to the wall. They seem to have grown outward precisely as does hoar frost in loose soil, where the moisture, rising by capillarity, freezes as soon as a certain level is reached, so that the older and first formed portions are ever pushed upward so long as the supply below is continued. As in the formation of hoar frost, particles of earth are lifted upon the tops of the ice spicules, so here the growing gypsum having begun forming in a crevice not infrequently forces off pieces of the limestone of considerable size. . . . . As the crystals form and are pushed outward they are in most cases in a condition of strain, which causes them to curl and twist in a remarkable manner, as shown.[22]

[21] *Op. cit.*, p. 151.

[22] *Op. cit.*, p. 81.

J. M. Weller (1927) disagreed with this concept of formation under strain. He states:

> It has previously been supposed that the curvature of these bundles of gypsum fibers has been produced in some way by pressure, perhaps due to growth at the base, but microscopic examination of thin sections of the gypsum from Great Crystal Cave shows that this is not the case. Under crossed nicols the gypsum mass may be seen to be composed of a group of fibers which, in their relations to one another, suggest twinning. In a single field these fibers curve through an angle of 40 degrees and yet each fiber extinguishes practically simultaneously throughout its whole length, proving conclusively that although the fiber is curved the crystallographic axes are straight and, therefore, the curving of these aggregates has been produced by the original manner of growth and is not due to any sort of force, pressure or deformation. Why the gypsum should have grown in this manner is not understood.[23]

The present investigation supports Weller's conclusion. The basis for the original theory of curvature by strain is probably based upon the flexible nature of gypsum. The sodium thiosulphate used to form the artificial rosettes is not flexible like gypsum and would break rather than bend if strained. Since the crystals did not break, it is evident that they were not strained.

It is not necessary to postulate strain, because the curve can be explained entirely by the differences in the growth rate of the opposite sides of the petals. The side that grows the faster should be the side toward the source of supply of the solution. In both the natural gypsum flowers and the artificial rosettes the curvature is away from the center of the cluster. The center always corresponds to the source of supply of solution for the artificial rosettes. Often, in both the natural and the artificial rosettes planes of darker deposit pass perpendicularly through the mass of fibers. These planes are evidently caused by a change in the growth rate. The planes are relatively flat, which indicates that adjacent crystals grow at nearly equal rates. Different planes in the same mass are not parallel; they are farther apart nearer the center of the rosette. This is another indication that the crystals grow more rapidly near the source of supply.

[23] "The Geology of Edmonson County," *Ky. Geol. Surv., Geol. Repts.*, Vol. XXVIII (6th ser., 1927), pp. 70–71.

It is difficult to visualize deposition taking place in the pores of the rock because deposition depends upon evaporation and air must reach the solution before evaporation can occur. This means that the air must circulate in rock at least as far as the bottom of the pores from which the formations are growing. There is good evidence indicating that it can penetrate even farther. In places in Great Onyx Cave, Kentucky, satin spar bands and single crystals of gypsum are found in the limestone as much as 3 inches from the surface. These crystals not only succeed in growing in the limestone, but they exert enough pressure to cause considerable disruption and spalling of the rock. This affords an excellent illustration of the pressure which can be produced by growing crystals.

The composition of stalactites, stalagmites, helictites, and most cavern formations reflects the composition of the ground water. In helictites, where the solution evaporates entirely, the relative proportions of the minerals will be the same as in the ground water. Gypsum flowers differ from the other formations in this respect, for they are composed of practically pure gypsum. It would be unusual to find that the solutions supplying them contained nothing but gypsum. A more logical explanation is that the gypsum was produced by a partial evaporation and that the solutions went elsewhere to deposit the remainder of their load.

Since the gypsum is deposited in the pores of the rock, it seems reasonable to assume that this is also true of the other mineral components of the ground water. The porosity of the Mammoth Cave and similar limestones is enough to permit a considerable addition of mineral matter. The addition of calcium carbonate—the chief mineral component—would not be readily recognized because it takes place as slight additions to innumerable small crystals already present in the rock. Its addition might possibly be detected by comparing the porosity of the surface rock and of the deeper, unaltered rock.

SUMMARY OF CONCLUSIONS

If it can be assumed that the natural helictites are formed by the same process as the artificial ones, a number of controversial problems can be solved. These experiments indicate that helictites grow

at the free end and not at the attached end. They also indicate that helictites are tubes that assume their crooked shape as a result of the chance orientation of the crystals at the growing end of the tube. Helictites, instead of stalactites, are produced when there is a hydrostatic head of solution and when there is a very small flow. In helictites this flow is confined to the channel at the center of the formation.

Most natural helictites are composed of calcite, but some aragonite ones are known. The presence of sulphate in those composed of aragonite suggests that it may have caused the precipitation of aragonite instead of calcite.

Experiments in growing formations similar to gypsum flowers indicate that porous rock is necessary for the growth of the natural flowers. The solution evaporates in the pores to produce a deposit which is forced outward, by more deposit forming below it, to form acicular crystals. The gypsum flowers grow at the attached end and not at the free end as helictites do. The curve of the petals of a gypsum flower is not caused by any straining of the crystal but by a more rapid growth on one side than on the other. The curvature of the petals is generally away from the center of the flower because the solution is supplied more rapidly to the center than to the sides.

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THE BOGOTA FAULT, COLOMBIA, SOUTH AMERICA[1]

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ABSTRACT

The abrupt scarp forming the eastern boundary of the Sabana de Bogotá in Colombia appears to be of considerable geologic significance. Detailed mapping up the face of Monserrate, immediately east of the city of Bogotá, discloses the fact that Cretaceous rocks overlie probable Tertiary beds, with a crushed zone of some 15 meters in thickness at the contact. It is the opinion of the writer that the scarp is the direct result of overthrusting, with differential erosion having played some part in its further development. The feature is here named the Monserrate-Guadalupe scarp, and the overthrust, the Bogotá fault.

INTRODUCTION

The great Cordillera de los Andes divides into three distinct ranges at the Colombian-Ecuadorian border, and these ranges continue northerly through Colombia, where they are known as the Eastern, Central, and Western Cordilleras, separated by the Magdalena and the Cauca valleys, respectively. Bogotá, the capital city of Colombia, is situated on a broad, flat grassland (the Sabana de Bogotá), high in the Eastern Cordillera, at an elevation of approximately 2,600 meters (8,500 feet) above sea-level. The Sabana is completely surrounded by mountains, which rise another 700–1,000 meters (2,200–3,200 feet) above its general level. The mountains at whose base the city lies form the crest of the Eastern Cordillera, and, incidentally, the eastern margin of the Sabana.

A considerable amount of geologic work has been done in Colombia, both of a regional and of a local character, although the geology of much of the country is still but imperfectly known. Prominent among the men who have contributed to the knowledge of the geology of the Eastern Cordillera are Karsten, Hettner, Stille, Scheibe, Anderson, Stutzer, Grosse, Hubach, Alvarado,[2] and others, many of whom were in the employ of the Colombian government, and some

[1] Published by permission of the Director, Department of Petroleum, Ministry of National Economy.

[2] See bibliography.

of whose work is still awaiting publication. In addition, a large quantity of geologic data has been accumulated by the various oil companies active in Colombia for the last twenty years, but, unfortunately, little has been published.

So far as could be ascertained, no detailed geologic work had been done in the immediate vicinity of Bogotá prior to that here recorded. The present writer, while awaiting the assembling of a field party for other geologic work, took advantage of the excellent topographic

FIG. 1.—Monserrate, part of the Monserrate-Guadalupe scarp, from the National Library, Bogotá. The deep notch to the right is the canyon of Rio San Francisco.

maps of Bogotá prepared by the Instituto Geográphico Militar on the occasion of the city's fourth centennial in 1938 to do detailed mapping across the scarp that lies just east of the city. The data recorded at this time, together with observations made later on a reconnaissance across the Eastern Cordillera, form the basis of this report. Most of the field work was done in March and April, 1939.

Acknowledgment is made of the splendid co-operation of all government officials concerned. Dr. Gabriel Gómez, Dr. Benjamín Alvarado, and Dr. Felix Mendoza, respectively director, chief geologist, and chief petroleum engineer of the Department of Petroleum, gave unstintingly of their time and friendly criticism. Without the approval of Dr. Jorge Gartner, Minister of National Economy, and

Dr. Eduardo Santos, President of the Republic of Colombia, the work could not have been undertaken. The writer was accompanied in the field during the time spent upon this particular problem by Dr. Eduardo Chaves and Mr. Luís Camargo. Officials and geologists of the Richmond, Shell, Socony-Vacuum, Texas, and Tropical oil companies were exceedingly kind in their offers of any information available.

FIG. 2.—Scarp of the Bogotá fault some ten miles south of the city, from the Bogotá-Villavicencio highway.

GEOLOGY

One of the most striking features in the vicinity of Bogotá is the great scarp at whose base the city is situated. This scarp rises abruptly from the general level of the Sabana and attains elevations of 3,200 meters (10,400 feet) in Monserrate, 3,280 meters (10,660 feet) in Guadalupe, and more than 3,600 meters (11,700 feet) in the peaks farther south. The scarp forms the eastern boundary of the Sabana for its entire length, a distance of some tens of kilometers, and it continues to the south of Bogotá for an unknown distance. Where crossed by the Bogotá-Villavicencio highway 16 kilometers (10 miles) south of the city, it is still very prominent. From its size and extent it appears to be a major feature, and therefore its nature

is of considerable significance in connection with the geology of the Eastern Cordillera as a whole.

Several possibilities suggest themselves as to the origin of the scarp. It may be simply the result of differential erosion of the two distinct types of sediments involved. It is, perhaps, the result of a combination of folding and differential erosion. Or it may be a fault scarp modified to some degree by erosion. These possibilities will be considered.

STRATIGRAPHY

Guadalupe formation.—The oldest rocks exposed in the mountains in the immediate vicinity of Bogotá are those of the Guadalupe formation, for which the type locality is Guadalupe Mountain, a peak just east of the city and south of Monserrate. The formation is widespread in the Eastern Cordillera, and not only comprises the backbone of the eastern rim of the Sabana but also forms the western frontal scarp of the range as a whole.

The predominant rock type in the Guadalupe formation is sandstone. This is generally thick-bedded, with individual strata up to 2 meters (6 feet) in thickness, and is generally yellowish in color, although it may be white, dark gray, or brown. The sandstone is fine to medium in grain, and is composed of subrounded particles of quartz with variable amounts of kaolinite, cemented in some instances by iron oxide, and in others by silica. In the former case, the sandstone is rather soft and friable, and is a source for sand used in the making of plaster and concrete in the city; in the other case, it is hard and quartzitic.

Shale occurs as partings between the sandstone beds, and occasionally becomes rather prominent, although always remaining subordinate to the sandstone. The shale as a rule is soft, thinly laminated, gray in color, and frequently micaceous, but in places is found as a hard, bluish-gray, siliceous type.

The beds of the Guadalupe formation dip in general to the east in the locality under discussion. The dips are less steep up the face of Monserrate than in the canyon of Rio San Francisco, which has cut a deep notch in the scarp almost at right angles to the strike of the beds, separating Monserrate from Guadalupe Mountain. Up the

scarp, recorded dips vary from 15° to 40° to the east; in the canyon, from 53° to 78°.

On the crest of Monserrate, and also along the face of Guadalupe, reverse dips were noted. It was the opinion of the writer at first that these reverse dips were the result of overturning to the east; the evidence, however, suggests that the westerly-dipping beds are part of a large fold, presumably involved in the faulting, and now largely removed by erosion. That the formation as a whole is not overturned is suggested by the occurrence of a few fossil marine pelecypods with the shells concave downward in the sandstones near the base of the formation on Monserrate.

In the particular area being considered, the Guadalupe formation overlies the Guaduas formation, which is next to be described. This is an abnormal relationship as is indicated by the fact that on the western side of the Sabana, where the beds are almost flat lying, the Guadalupe is overlain by the Guaduas, and in turn overlies the Villeta formation (Middle Cretaceous), in both cases with normal, depositional contacts.

The age of the Guadalupe formation is generally accepted as Upper Cretaceous, although the type locality has proved so far to be barren of fossils. The few marine fossils occurring in the sandstones forming the face of Monserrate indicate a marine origin but are not diagnostic. Collections made by the writer from the continuation of the same series a few miles to the south, but stratigraphically lower in the section, have been determined as indicating a Cenomanian age by Dr. José Royo y Gómez, paleontologist in charge of the Geological Museum and Laboratory of the Ministry of National Economy.[3] Some poorly preserved specimens from higher in the section, and in beds more nearly the equivalents of those under discussion, have yet to be studied and reported upon, but on brief inspection suggest nothing out of keeping with the foregoing determination. Thus there seems to be no reason to question the accepted age of the Guadalupe formation as Upper Cretaceous.

Guaduas formation.—Rocks that have been assigned to the Guaduas formation outcrop along the foot of the Monserrate-Guadalupe

[3] José Royo y Gómez, "Informe provisional sobre los yacimientos fosilíferos de Bogotá a Villavicencio" (unpublished report, 1939; rev., 1940).

ridge in the vicinity of Bogotá and continue to the south for a distance of several miles. The type locality for the formation is far down the western slope of the Eastern Cordillera, near the town of Guaduas, the name apparently first having been applied by Hettner.[4] The formation is rather widely distributed throughout the Eastern Cordillera, where it is of considerable economic importance, since it is the source of most of the coal used in Colombia for both industrial and household purposes.

The principal lithologic types represented in the region here discussed are shale and sandstone, with the former predominating, particularly as the formation is followed to the south. The shales are variable in color, ranging through gray, yellow, green, red, maroon, and purple, and as a rule are characteristically mottled. They are soft and generally massive, and are much used in and around Bogotá for brick-making.

The sandstones in general are mottled yellow and red, rather soft, and in many places cross-bedded. Occasional hard beds weather out in sharp, wall-like ridges. The only coal observed by the writer in the vicinity of Bogotá occurs associated with the sandstone. This is up to a meter in thickness but is of rather poor quality, although it has been exploited to some extent in the past. The coal noted above as occurring in the Guaduas in other parts of the Eastern Cordillera is of good quality.

The dips in the Guaduas formation along the foot of Monserrate are in general easterly, like those in the overlying Guadalupe, but as a rule are steeper, ranging from 40° up to vertical. As the contact between the two formations is approached, there is a reversal in dip, suggesting drag folding. Some $2\frac{1}{2}$ miles south of Monserrate, in the canyon of Rio San Cristobal, at a distance of perhaps 1 mile west of the scarp, the Guaduas beds dip to the west. The dips steepen rapidly in the direction of the scarp, and still farther south, where the highway approaches the contact more closely, the dips again become easterly.

The above-mentioned cross-bedding in the sandstones, the lack of lamination in much of the shale, and the presence of coal, all sug-

[4] A. Hettner, "Die Kordillere von Bogotá, *Petermanns Mitteilungen*, Vol. XXII (1892), p. 16.

gest a nonmarine origin for the Guaduas. No fossils were found by the writer nor, to his knowledge, have any fossils other than plant remains been encountered in the formation.

Hettner[5] considered the formation at the type locality to be Cretaceous. More recently Butler[6] has stated that the Guaduas includes strata ranging in age from uppermost Cretaceous through the Eocene, and has divided the formation into three parts on the basis of lithology. These are a lower member containing coal, in which the predominant lithologic type is gray shale; a middle member in which sandstone and fine conglomerate predominate; and an upper member consisting largely of reddish and mottled clays and shales. The rocks in the vicinity of Bogotá appear to correspond more or less with the uppermost part of Butler's lower division, and in part, at least, to his middle division. Farther south the beds in contact with the Guadalupe probably correspond more closely with the upper division, which suggests that the contact cuts stratigraphically across the Guaduas. The present writer considers that the Guaduas formation of the Monserrate-Guadalupe ridge is of probable Eocene age.

Pleistocene terrace material.—Mantling much of the more gentle slope at the foot of the scarp proper is a blanket of unconsolidated, poorly sorted terrace material. The larger fragments are angular, and up to 2 meters (6 feet) in maximum diameter, many of them slickensided. The material appears to have been derived almost entirely from the Guadalupe formation, and to have been deposited in a series of alluvial fans or as an alluvial apron. A rude bedding is rather infrequently observed; at the foot of Monserrate the beds dip easterly at an angle of approximately 7°. The material, which is being dissected by present-day streams, is without question Pleistocene in age.

STRUCTURAL RELATIONSHIPS

It is evident from the foregoing that an abnormal relationship exists between the Guadalupe and the Guaduas formations, younger rocks being overlain by older. In order to determine the true cause of

[5] *Ibid.*

[6] John W. Butler, Jr., "Geology of Middle and Upper Magdalena Valley," *World Petroleum*, Vol. X (1939), pp. 95–100.

this, it is necessary to learn something of the nature of the contact between the two formations. Unfortunately, the contact is masked in most localities by the overburden of terrace material, making direct observation difficult. However, in a small canyon cutting the face of Monserrate, and along which the accompanying section was made, an excellent exposure exists.

The Guaduas formation, here consisting of rather hard sandstone with partings of slightly plastic shale, exhibits the usual easterly dip

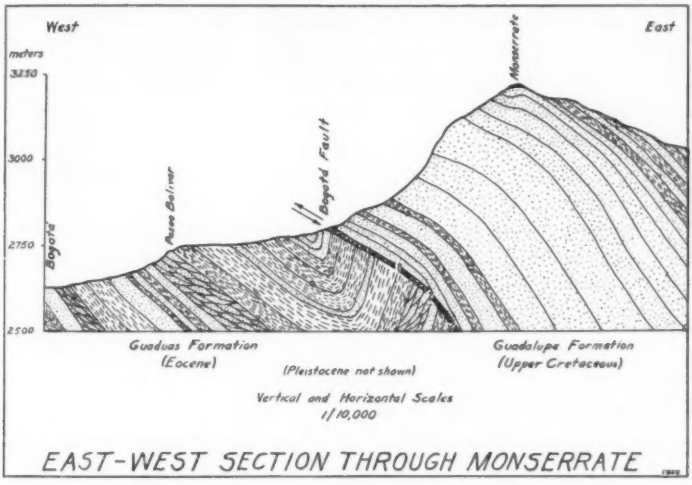

FIG. 3

below the contact, but, as the latter is approached, the dip changes to northwesterly at a high angle. Just upstream from this point occurs the easterly-dipping terrace material.

The contact is a zone of approximately 50 feet in thickness, consisting of sandstone so crushed and broken that all semblance of bedding has been destroyed. The material resembles aplite in outward appearance, although under the hand lens the sedimentary character of the original rock is still visible. Two seams of gouge, 15 and 25 centimeters (6 and 10 inches) thick respectively, occur in the zone, dipping at 30° slightly south of east. Slickensided sand-

stone, almost quartzite in character, occurs adjacent to the gouge, and pieces of sandstone are squeezed into lens shape.

Above the contact the rocks are less broken, and dips of 40° in an easterly direction are found in a rather massive sandstone. These dips become more gentle farther up the canyon, until normal dips of 20°–25° predominate.

There can be no doubt that the contact is a fault, and this is further substantiated by similar evidence where it is crossed by the highway farther to the south, and by the fact that the contact cuts stratigraphically across the Guaduas formation, as brought out earlier in the paper. From the low angle, the reverse sequence of beds, the drag fold in the Guaduas strata, and the previously postulated fold in the Guadalupe, it is obvious that the fault is an overthrust.

The amount of movement on the fault is unknown, and a more detailed knowledge of the contact as a whole and of the two formations is requisite for anything like an accurate determination. Butler[7] suggests a thickness for the Guaduas formation of more than 2,000 meters (6,500 feet), although whether or not this figure would apply to this particular locality it is impossible to say. The thickness of the Guadalupe formation exposed here is in excess of 600 meters (2,000 feet), although neither top nor bottom is exposed. This gives an approximate measure of the stratigraphic throw as 2,600 meters (8,500 feet). The heave also may be measurable in a few thousands of feet.

The time of movement on the fault is difficult to fix, further than post-Eocene. Seismic activity in the vicinity of Bogotá, which has been prevalent since earliest historic times,[8] is probably attributable to this fault, and indicates that movement is still continuing. The tilting of the terrace material near the contact is attributed by the writer to sagging of the relatively weak Guaduas rocks under the load of the overthrust Guadalupe, carrying the terrace material with it, and causing it to dip toward the scarp. This also indicates comparatively recent activity on the fault, although the presence of

[7] *Ibid.*

[8] J. Emilio Ramirez, S.J., "Earthquake History of Colombia," *Bull. Seism. Soc. Amer.*, Vol. XXIII (1933), pp. 13–22.

slickensided boulders of Guadalupe sandstone in the terrace material can be accounted for only by activity during or prior to the deposition of the latter. Furthermore, erosion has removed much of the evidence of folding in the Guadalupe formation from Monserrate, which suggests that the movement had its inception in the not too recent past. It is possible that the fault may be dated as early Pleistocene.

CONCLUSIONS

Based upon the evidence brought out in the foregoing, it is the opinion of the writer that the scarp forming the eastern margin of the Sabana de Bogotá, and here designated the Monserrate-Guadalupe scarp, is the result of an overthrust fault. Differential erosion has modified the scarp to some extent. It is the further opinion of the writer that this overthrust is still active and that it is a major structural feature of the Eastern Cordillera of Colombia.

BIBLIOGRAPHY

NOTE.—This bibliography is not exhaustive, and contains only references to the Eastern Cordillera of Colombia.

ANDERSON, F. M. "Original Source of Oil in Colombia," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. X (1926), pp. 382–404.

GROSSE, E. Numerous papers in *Compilación de los estudios geológicos oficiales en Colombia*, Vol. III (1935).

HETTNER, A. "Die Kordillere von Bogotá," *Petermanns Mitteilungen*, Vol. XXII, No. 104 (1892).

HUBACH, E. "Exploración en la región de Apulo—San Antonio—Viotá," *Boletín de minas y petróleo*, Vol. IV (1931).

KARSTEN, H. *Geologie de l'ancienne Colombie Bolivarienne*. Paris, 1886.

SCHEIBE, R. Numerous papers in *Compilación de los estudios geológicos oficiales en Colombia*, Vol. I (1933).

STILLE, H. "Geologische Studien im Gebiete des Rio Magdalena," 1907. Spanish translation in *Compilación de los estudios geológicos oficiales en Colombia*, Vol. IV (1938), pp. 125–82.

STUTZER, O. "Contribución a la geología de la Cordillera Oriental, regiones cerca de Bogotá," *Compilación de los estudios geológicos oficiales en Colombia*, Vol. II (1934), pp. 141–82.

REVIEWS

Manuel de paléontologie animale. By LÉON MORET. Paris: Masson & Cie, 1940. Pp. 675; figs. 241; pls. 12. $3.65.

Moret's *Manual of Animal Paleontology* is the most complete and readily teachable French work on the subject which has appeared since 1895, when Félex Bernard's epoch-making, 1,168-page *Eléménts de Paléontologie* was issued.

The book in essence comprises the outline of the course in paleontology taught for a number of years by Professor Moret at the University of Grenoble. Thus the work has, to a certain extent, stood the acid test of actual use at least in pre-publication form. The author apparently feels that the volume is more or less of elementary type, because, through it, he expects that students will be introduced to the subject. But he is also hopeful, and with reason, that specialists and amateur collectors will find the work of some value. For this reason he has devoted about three-quarters of the book to the Invertebrates, since they are more likely to be encountered by the average collector and are of greater utilitarian value in geology generally. The Vertebrates, by contrast, are discussed in a very condensed style, their story being used chiefly to throw light on the phenomena of evolution.

In addition to systematic descriptions of most of the major groups of animals represented in the fossil record, the author has, in practically all cases, compared the fossils with their living end-products and has gone into the ecological features of the forms involved, as well as their role in the phenomena of sedimentation. The valuable 30-page Introduction expresses clearly a number of the principles of paleontology, explains the processes of fossilization, makes reference to the general problems of evolution, and discusses the part that paleontology can play in their solution. Throughout the volume Professor Moret has attempted to depart from the scheme of earlier works by merging a systematic descriptive style with a paleobiologic approach. The reviewer believes that a generally creditable result has been achieved.

Altogether, more than 2,000 illustrations of uniform style are represented in the 241 figures. In all cases these are fine-line drawings which the author himself has copied and simplified from many sources, in the

main well chosen. There is also an alphabetical index comprising approximately 3,000 names of genera, sub-genera, families, and so forth. Thus the *Manual* is easy to handle and convenient to consult.

In spite of the fact that Moret's work has many fine features in its favor and is especially encouraging for the future of French paleontology (whose immediate past has not been too brilliant), there are numerous improvements which might well be made in a second edition. Many experts, for instance, will find much to condemn concerning some of the classifications employed. Such criticisms, of course, are inescapable, since they are directed at all works on paleontology. But practically all students will regard the amount of space devoted to the various groups as especially spotty. For example, the Foraminifera receive a more extended treatment (67 pages) than in any other general paleontology, whereas the discussion of the starfish and ophiuroids together is almost unaccountably compressed into a half-page. Holothurians are given even less space, which is equally surprising in view of the fact that most of the work on fossils of this group probably has been done by French workers.

In conclusion, it is probably fair to state that Moret's *Manual* possesses a number of the good features of both Twenhofel and Shrock's *Invertebrate Paleontology* and Swinnerton's *Outlines of Paleontology*. Its Invertebrate section is, however, not so detailed or so well documented as in the American volume, nor is its philosophy as clean cut and as succinctly stated as that expressed throughout the English *Outlines*.

CAREY CRONEIS

Geologic Map and Guide of Oahu, Hawaii. By H. T. STEARNS. Honolulu: Division of Hydrography of the Territory of Hawaii in co-operation with the Geological Survey, U.S. Department of the Interior, 1939. Pp. 75; pls. 6; figs. 23.

This bulletin is one of four published by the Division of Hydrography of the Territory of Hawaii, all of which relate to the geology and groundwater resources of the island of Oahu. In Bulletin No. 2 is presented a detailed geologic map of Oahu, as well as a guide to the geology along the main highways which can be used advantageously in connection with the map. The text also includes descriptions of the cross sections given on the map. A sketch showing principal roads and geologic features is printed on the inside of the cover. The diagrams illustrating the formation of the island and photographs of outcrops are unusually good.

LOUISE BARTON FREEMAN

The Pre-Cambrian—Cambrian Succession: The General and Economic Geology of These Systems, in Portions of South Australia. By RALPH W. SEGNIT. (Geological Survey of South Australia, Department of Mines, Bull. 18.) Adelaide, Australia: Frank Trigg (government printer), 1939. Pp. 191; figs. 78.

This bulletin deals with a new classification of the Middle and Upper pre-Cambrian and Lower Cambrian formations in certain portions of southern Australia and with the geology of some of the best-known mining fields. The book is divided into thirteen parts, the last of which gives the summary and conclusions. Each of the first twelve parts describes in some detail the stratigraphy, structure, and mineral deposits of a particular section. Such systematic geological mapping of considerable areas has made it possible for the author to establish the full sequence from the base of the Sturtian tillite to the limestones containing definite Cambrian fossils and to indicate the character and attitude of the formations underlying and overlying these uppermost pre-Cambrian rocks.

Each part is illustrated with a geological map and numerous photographs and diagrams.

LOUISE BARTON FREEMAN